Structure of the Canadian Appalachians

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ABSTRACT

Precambrian deformational events in the Canadian Appalachians are (1) Middle Proterozoic polyphase structures associated with high grade metamorphism in Grenvillian inliers and within the Avalon Zone in northwestern Cape Breton Island and eastern Cobequid Highlands; and (2) *Avalonian*, Late Proterozoic-Cambrian single phase to polyphase structures accompanied by low to high grades of metamorphism in the Avalon Zone. The sporadic *Avalonian* structures are inferred to have developed in a rifted magmatic arc complex in a transpressional environment that produced synchronous intra-arc rifting, transcurrent faulting and compressional deformation.

Paleozoic deformational events in the Canadian Appalachians correspond to narrow diachronous events in the Ordovician, Silurian, and early-middle Devonian, whereas late Devonian, Carboniferous and Permian deformational events are widespread and braodly synchronous. Along the western side of the Canadian Appalachians, the *Taconian* deformational event starts diachronously throughout the Ordovician and corresponds to the N-NW accretion of the Notre Dame, Ascot-Weedon, St. Victor and various ophiolitic massifs (volcanic arc and peri-arc terranes) over cratonic North America. Within the eastern half of the Central Mobile Belt, the late Cambrian-Early Ordovician, *Penobscotian* deformational event corresponds to the ?SE-accretion of the Exploits composite terrane (various volcanic arc and peri-arc terranes) over the Gander terrane (?continental rise). In the centre of the orogen, the late Ordovician-Silurian, *Beothukan* deformational event corresponds to the SE-accretion of the Notre Dame over the Exploits-Gander composite terranes. Along the southeastern side of the Central Mobile Belt, the Silurian *Ganderian* deformation event corresponds to the N-NE, sinistral transcurrent accretion of the Avalon (continetal terrane) over the Gander-Exploits composite terranes. Along the southeastern half of the orogen, the early-middle Devonian *Acadian* deformation event corresponds to the W-accretion of the Meguma terrane (intramontane basin or continetal rise) over the Avalon composite terrane. Affecting the entire orogen, the late Devonian, Carboniferous and Permian, *Acadian-Alleghanian* deformational events corresponds to the E-W convergence between Laurentia and Gondwana (continent-continent collision). These Paleozoic deformational events together constitute the Appalachian Orogeny.

The Late Proterozoic and Paleozoic deformational events in the Canadian Appalachians are explicable in terms of plate tectonics. Plate tectonic theory predicts that most deformation is associated with subduction and terrane accretion, with some deformation may be associated with transform/transcurrent movements. Deformation associated with subduction varies between two end members: (1) where the tectonic regime is dominated by subduction of oceanic lithosphere containing small terranes, a *narrow* surface zone of accretionary deformation along the subduction zone which starts *diachronously* on the subducting plate at the trench as material is transferred from the subducting plate to the over-riding plate; and (2) where continent-continent collision is occurring, a *wide* surface zone of accretionary deformation that starts *synchronously* or with limited diachronism.

STRUCTURE OF THE CANADIAN APPALACHIANS

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INTRODUCTION

The Canadian Appalachians is generally defined to include those areas deformed by Paleozoic structures (Fig. 1). This includes Newfoundland, Nova Scotia, New Brunswick and Quebec south of the St. Lawrence River. The Appalachian structural front closely follows the St. Lawrence River in Quebec passing into the the Gulf of St. Lawrence and the Strait of Belle Isle between Newfoundland and Quebec. Along most of its length, it lies under water, however, it is exposed in southern Quebec, where it has been defined as the western limit of thrusting (St. Julien and Hubert, 1975). West of this thrust front, the Paleozoic rocks are gently folded (and could be included in the Appalachian Orogen) passing northwestwards into undeformed, gently dipping rocks. In some places, this structural front follows the Paleozoic-Precambrian boundary, but in other places it traverses Paleozoic miogeoclinal rocks deposited on the Precambrian North American craton.

In plan view, the shape of the Appalachian Orogen in Canada is elongate, sinuous and Z-shaped as it follows the Quebec Salient, the St. Lawrence Recess, and the Newfoundland Salient (Thomas, 1977: Fig. 1). This shape is inherited from the original irregular edge of the North American craton: reentrant-promontory-reentrant, respectively. In general, the major structures also parallel this sinuous trend in the orogen (Fig. 1), suggesting that their surface traces are more a function of the initial geometry of the North American margin than of kinematics. A similar, sigmoidal, Z-shaped trend of the structures is also present in the Dunnage and Gander Zones in Newfoundland, but here their geometry is attributable to the sinistral accretion of the Avalon Zone with its irregular northwestern margin. The intensity of Ordovician deformation south of the Appalachian structural front passes from a zone of single phase folding and thrusting generally associated with low grade metamorphism into a zone of complex polyphase deformation usually accompanied by higher grades of metamorphism in the internal parts of the orogen (Keppie *et al.*, 1982). In general, Paleozoic deformation records the accretion between the zones and subzones in the Appalachians which usually started earlier in the internal parts of the orogen and its onset became progressively younger towards the margins. High grade metamorphism is associated with two Early Paleozoic accretionary prisms bounding the Dunnage Zone. The Devonian - Early Carboniferous rocks are most deformed in NE-SW zones passing through southern New Brunswick, Nova Scotia and from St. Georges Bay to White Bay in Newfoundland. Middle-late Devonian rocks are thrust N-NW in the Gaspe and southern Quebec (Keppie *et al.*, 1982). In contrast, the Late Carboniferous - Permian rocks are generally little deformed or undeformed.

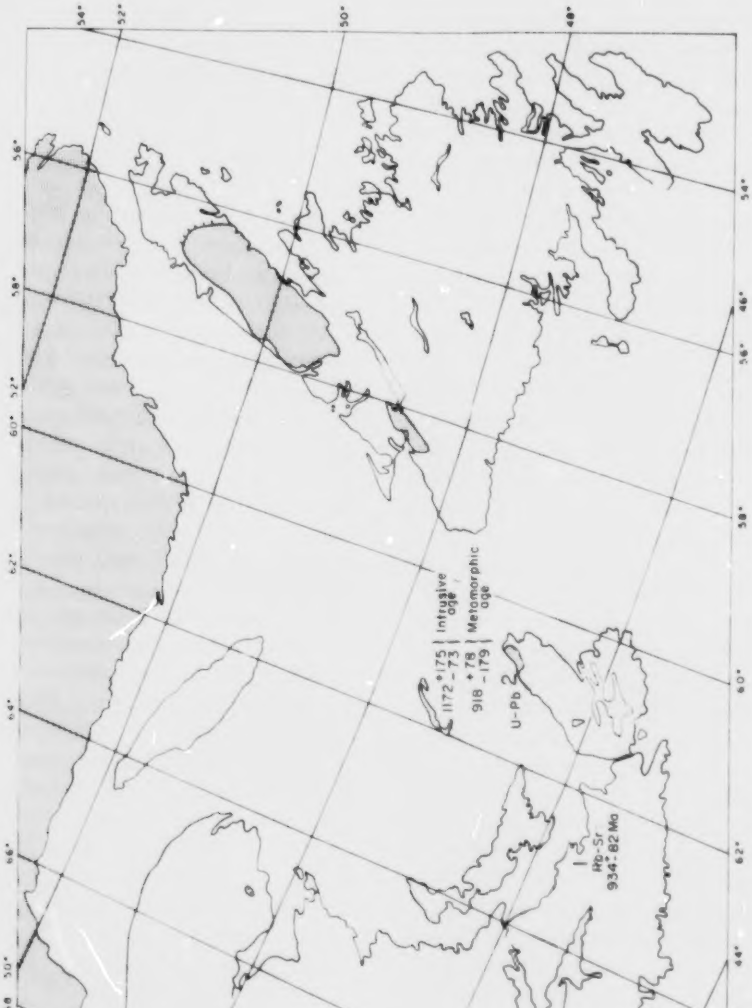
The geometry of the orogen in cross-section (Fig. 1) is constrained by seismic reflection data in the Quebec-Gulf of Maine transect, by various lines in the Gulf of St. Lawrence - Cabot Strait and in the Newfoundland by a transect just north of the island (St. Julien *et al.*, 1983; Roksandic and Granger, 1981; Keen *et al.*, 1986; Stewart *et al.*, 1986; Marillier *et al.*, 1989; Keen *et al.*, 1991). In Newfoundland, the structure appears to consist of a series of allochthons that diverge from an axis in the Dunnage Zone. These allochthons are sharply truncated by a vertical fault boundary (Dover-Hermitage Faults) along the northwestern margin of the Avalon Zone. The structure in the Avalon Zone of Newfoundland consists of gently folded horsts and grabens. This contrasts markedly with the complex structure of the Gander and Dunnage Zones. On the mainland, a similar structure with opposing vergence across a central axis in the Dunnage Zone has been modified by a series of thrusts verging mainly towards the ancient North American craton. In contrast to Newfoundland, the northern boundary of the Avalon Zone on the mainland appears to be a southeasterly dipping, listric fault. The rocks of the Avalon Zone on the mainland have been more involved in Paleozoic deformation than those in Newfoundland. The Meguma Zone also appears to be an allochthon thrust westwards over the Avalon Zone (Keppie and Dallmeyer, 1987; Keen *et al.*, 1991) and bounded on its northern margin by a dextral shear zone (Minas Fault). Listric normal faulting produced half grabens of Mesozoic age, such as the Bay of Fundy, and appears to have reactivated some pre-existing thrusts.

Some workers in the Canadian Appalachians have envisaged deformation in terms relatively short events that could be widely correlated (e.g. Kennedy and St. Julien in Williams *et al.*, 1972). However, independent constraints upon the time of deformation were only available at a few selected sites, and temporal correlations were based only on similarity of structural styles. Furthermore, sequential structures were assigned to different phases of deformation or orogenies. Problems with such a scheme began to emerge in the compilation of the Structural Map of the Appalachian Orogen in Canada (Keppie *et al.*, 1982; 1983) when several deformational events fell outside the originally defined, short-lived events. Also the demonstration that polyphase structures may be produced in shear zones by one phase of progressive deformation suggests caution in assigning different times of deformation to sequential structures. The present synthesis preferentially uses sites where independent data exists for the age of structures. This analysis shows that the hypothesis of deformation being restricted to short-lived, but widespread events must be abandoned. In their place, long-lived, more geographically restricted, deformational events appear to characterize the early Paleozoic evolution of the Appalachians and may be related to terrane accretionary events. This was followed by a terminal, diachronous, long-lived event which affected the entire width of the Canadian Appalachians that may be related to continent-continent collision. These events may be named either by redefining older terms or assigning new names. As relatively little modification is required for the Taconian and Acadian-Alleghanian events, redefinition of these terms seems preferable. Precambrian deformation occurs in the Grenvillian inliers along the western margin of the orogen and in the Avalon Zone along the eastern side of the orogen where it varies considerably in style and intensity. The late Proterozoic deformation in the Avalon Zone is assigned to the Avalonian. Early Paleozoic deformation in the western part of the orogen is assigned to the Taconian deformational event and may be related to amalgamation of the Humber Zone and Notre Dame subzone. The term Penobscotian is restricted to deformational events associated with the amalgamation of the Exploits subzone and the Gander Zone. The term Beothukian is introduced for the Late Ordovician-Early Devonian deformation associated with the amalgamation of the Notre Dame and Exploits subzones. The term Ganderian is resurrected for Silurian deformation that is most intense in the Gander Zone and may be related to accretion of the Avalon Zone. Distinction between Beothukian and Ganderian becomes difficult where they merge. Late Paleozoic deformation across the entire orogen is assigned to the Acadian-Alleghanian. It is sometimes difficult to assign Devonian structures to either the terminal stages of these early Paleozoic accretionary events or to the Acadian and Alleghanian deformational events. The Paleozoic accretionary events may be grouped into the Appalachian Orogeny.

An earlier synthesis of the Canadian Appalachians (included a structural review, Kennedy and St. Julien in Williams *et al.*, 1972) was written when modern structural studies in the Canadian Appalachians were in their infancy. Although numerous structural studies have taken place in the intervening period, they tend to have concentrated on structural history and geometry in local, isolated areas and generally do not include kinematic or strain analyses. Since 1972, much regional mapping has taken place, and is still continuing, in many parts of the Canadian Appalachians. These studies formed the basis for a volume of synthetic structural papers (in St. Julien and Beland, 1982) and the Structural Map of the Appalachian Orogen in Canada (Keppie *et al.*, 1982), which depicts the age and spatial distribution of the structures supplemented by a brief summary on the times of deformation and their constraints (Keppie *et al.*, 1983). Two geological profiles across the Canadian Appalachians by Dewey *et al.*, (1983) and Rast (1983) contain sections on the structure. With the advent of plate tectonics, there have been many models for the development of the Canadian Appalachians Zones, which have progressed from simple subduction models (e.g. Bird and Dewey, 1970) to those involving the accretion of terranes (Keppie, 1989). In the simple subduction models, the origin of the structures depended upon the inferred polarity of subduction: back-arc or accretionary prism deformation, although some models invoke delamination and flake tectonics also. With the advent of terranes, it was proposed that deformation was produced collision and amalgamation of terranes (Keppie, 1985, 1989). These various hypotheses may be tested using the data in this chapter.

This chapter will proceed historically from oldest to youngest with the following sections:

Proterozoic-Cambrian: middle Proterozoic
late Proterozoic-Cambrian
Paleozoic: Taconian
Penobscotian
Beothukian
Ganderian
Acadian and Alleghanian
Mesozoic


CA. 900 Ma DEFORMATION
1-7 LOCALITIES IN ACCOMPANYING TABLE
GRENVILLIAN GNEISSES ≥ 900 Ma
NORMAL FAULT
THRUST FAULT
UNCONFORMITY
1172 +175 -73 Intrusive age
918 +78 -179 Metamorphic age
U-Pb 2
1 Rb-Sr 934±82 Ma
54°
52°
50°
48°
46°
44°
42°
56°
58°
60°
62°
64°
66°
68°
70°
72°
74°

AVALON COMPOSITE TERRANE		
LOCATION	1. MT. THOM	2. NORTHERNMOST CAPE BRETON ISLAND
Oldest post-tectonic rocks or event	Granite pluton (undated: presumed Carboniferous)	
Nature and age of tectonism, structure, metamorphism and stratigraphy	(Rb-Sr WRI = 934±82 Ma). Amphibolite facies metamorphism accompanied S_2 foliation parallel to axial planes of isoclinal folds. S_1 foliation.	(U/Pb zircon lower intercept from syenite = 918 + 78-179 Ma) dates high grade metamorphism and polyphase deformation.
Youngest pretectonic rocks	Mt. Thom Complex (age unknown)	Syenite (U/Pb upper intercept = 1172 + 175-73 Ma) intrudes Pollets Cove Brook group = (Pleasant Bay complex).

Figure 2. Map of middle Proterozoic structures (ca. 900 Ma) in and around the Appalachian Orogen, with a table listing the nature and contraints on the age of the tectonism. References by column number:(1) Donohoe, 1976; Donohoe and Cullen, 1983; Gaudette *et al.*, 1983, 1984; Doig *et al.*, 1989, 1991; Nance and Murphy, 1990. (2) Raeside *et al.*, 1986; Barr *et al.*, 1987; Keppie *et al.*, 1991.

Each section will contain a description of the structures and constraints on their age followed by an interpretation of their origin.

PROTEROZOIC-CAMBRIAN STRUCTURES

Proterozoic rocks outcrop (i) along the northwest flank of the Appalachians as the autochthonous - para-autochthonous Grenvillian basement to the Paleozoic miogeocline, (ii) as isolated basement blocks within the Dunnage Zone, and (iii) in the Avalon Zone. Structures in these Proterozoic rocks may be broadly grouped into ca. 900 Ma (middle Proterozoic) and late Proterozoic-Cambrian (Figs. 2 and 3). In general, the geometry, ages and kinematics of the ca. 900 Ma structures are poorly defined, allowing no coherent picture to be made. Although the geometry of the late Proterozoic structures is better known, age constraints and kinematics are poorly understood.

Middle Proterozoic Structures

HUMBER ZONE: GRENVILLIAN INLIERS

Middle Proterozoic, Grenvillian rocks form the cratonic basement along the north-western margin of the orogen and also occur as upthrust blocks generally within the Humber Zone. Structures in the isolated basement blocks within the Appalachian orogen are polyphase. In the external inliers, the plastic, polyphase structures have generally been assigned to the Grenvillian orogeny and only brittle east-dipping thrusts postdated by NW-SE and NE-SW steep faults are assigned to the Paleozoic (Erdmer, 1986). Some of these brittle structures are cut by the Devils Room pluton dated at 398 +27/-7 Ma (U-Pb zircon age; Erdmer, 1986). In contrast, internal inliers have been extensively affected by the Paleozoic metamorphism and deformation inducing transposition and making the distinction between Precambrian and Paleozoic structures difficult (DeWit, 1980).

AVALON ZONE

Gneisses outcrop at several locations in the Avalon Zone in the Cobequid Highlands and Cape Breton Island in Nova Scotia and some of the structures within them are considered to have formed in middle Proterozoic times (Fig. 2). How much of Cape Breton Island belongs to the Avalon Zone is controversial (e.g. Keppie, 1990; Barr and Raeside, 1990; Jamieson *et al.*, 1991; Keppie *et al.*, 1991; Keppie and Dallmeyer, 1991; Barr *et al.*, 1991). Some authors have correlated the gneisses in northwestern Cape Breton Island with autochthonous Grenvillian rocks of the Humber Zone, while others believe that these gneisses constitute the continental basement of the Avalon Zone. This subject is discussed in Keppie *et al.* (1991) and will not be reiterated here. In this chapter, all of Cape Breton Island will be considered part of the Avalon Zone.

In the eastern Cobequid Highlands, gneisses outcrop in the Mt. Thom Complex. Polyphase deformation in these ortho- and para-gneisses consists of an early foliation folded by isoclinal folds associated with an axial planar foliation and amphibolite facies metamorphism, in turn deformed by a crenulation cleavage (Donohoe, 1976; Donohoe and Cullen, 1983). A poorly defined, Rb-Sr whole rock ioschron on this complex yielded a 934 ± 82 Ma age (Gaudette *et al.*, 1983, 1984). This age may date the high grade metamorphic event although isotopic mixing and/or resetting by the last event cannot be ruled out and could account for the large error.

In the western Cape Breton Highlands, the Pleasant Bay Complex is made up of biotite gneiss, amphibolitic gneiss, gneissic granodiorite, marble, quartzite and schist cut by lit-par-lit gneiss (Currie, 1987). The gneisses display polyphase deformation and several episodes of metamorphism: the early metamorphism is largely obscured by a high pressure and high temperature upper amphibolite (-granulite) facies metamorphism overprinted by late retrograde metamorphism. The Lowland Cove syenite, intrusive into gneisses of the Pollets Cove Brook Group (= Pleasant Bay Complex) in northwestern Cape Breton Island, has yielded an U-Pb chord with lower intercept age of 918 +78/-179 Ma and an upper intercept age of 1172 +135/-73 Ma interpreted to reflect the ages of metamorphism/deformation and intrusion, respectively (Barr *et al.*, 1987; Keppie *et al.*, 1991).

Late Proterozoic-Cambrian Structures

Structures of Late Proterozoic-Cambrian age occur in the Avalon Zone (Figs. 3) where they have been attributed to the Avalonian Orogeny. Normal faulting and intrusion of dykes accompanying the

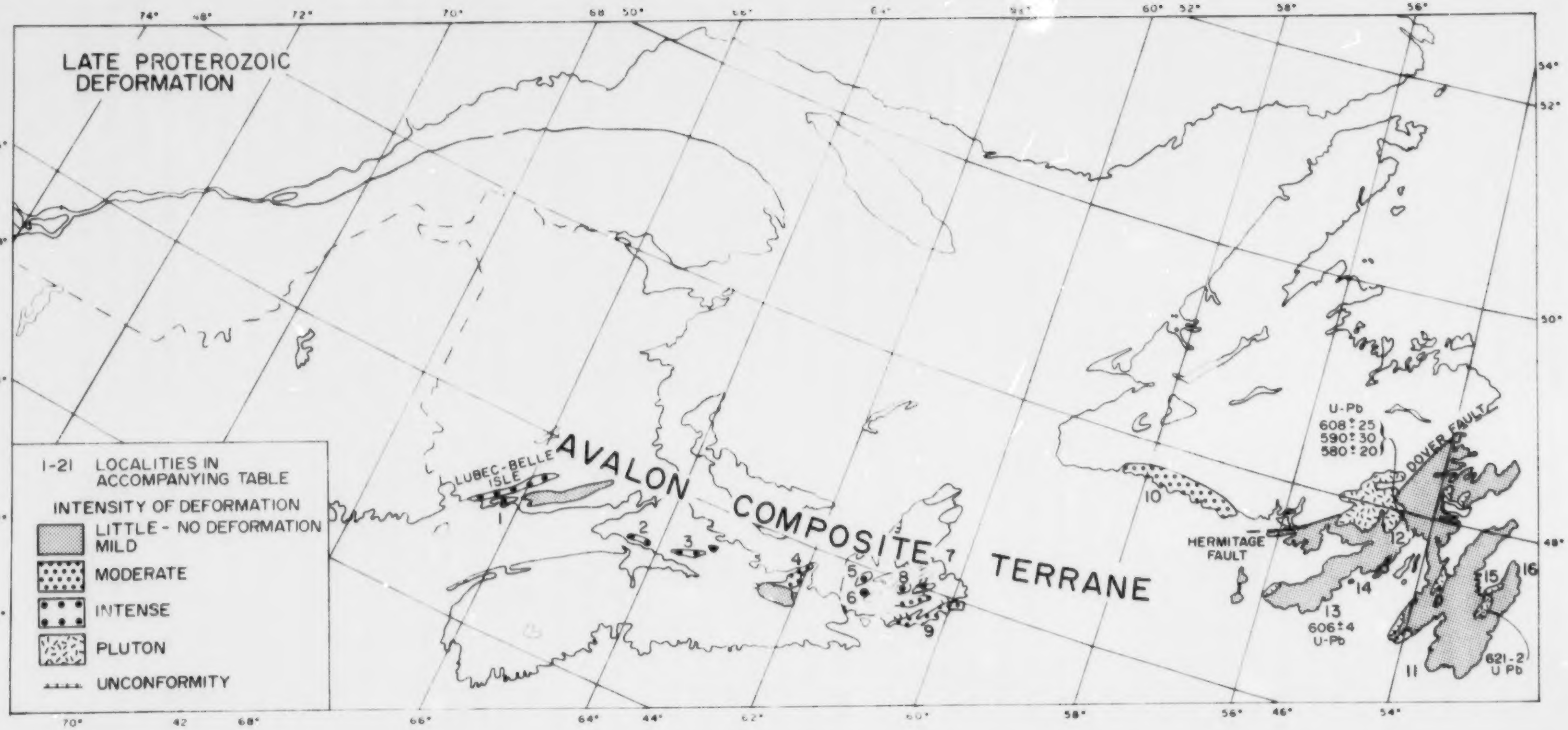

LATE PROTEROZOIC DEFORMATION
1-21 LOCALITIES IN ACCOMPANYING TABLE
INTENSITY OF DEFORMATION
LITTLE - NO DEFORMATION MILD
MODERATE
INTENSE
PLUTON
UNCONFORMITY
AVALON COMPOSITE TERRANE
LUBEC-BELLE ISLE
HERMITAGE FAULT
DOVER FAULT
U-Pb 608±25 590±30 580±20
606±4 U-Pb
621-2 U Pb
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
74°
48°
72°
70°
68
50°
66°
64°
62°
60°
52°
58°
56°
54°
46°
44°
42°

Figure 3. Map of late Proterozic-Cambrian structures with a table (in pocket) listing the nature and constraints on the age of the tectonism at localities marked on the map.

birth of Iapetus were taking place contemporaneously in the Humber Zone.

AVALON ZONE

Identification of late Proterozoic structures in the Avalon Zone is still fraught with uncertainties in many areas. This is largely due to a combination of factors including a dirth of critical contacts many of which are tectonic, a lack of isotopic data, and the difficulty in separating Paleozoic from Precambrian structures. Critical localities and summaries of the relationships are shown in Figure 3. The nature of the late Proterozoic deformation varies considerably from mild and nonpenetrative faulting, warping, disconformities and unconformities in Newfoundland and parts of southern New Brunswick, through moderate single phase to intense polyphase deformation in Nova Scotia and possibly some parts of southern New Brunswick. It appears that such inhomogeneous deformation is spatially related to shear zones and varies from moderate to intense in both steep and gently dipping shear zones but is absent in intervening areas.

There is still considerable divergence of opinion on the ages of the structures in the Precambrian rocks of the Nova Scotia and New Brunswick Avalon Zone. Thus, on the one hand, some workers (e.g. Wiebe, 1971; Rast, 1983; Jamieson, 1984; and Nance and Murphy, 1990) believed that there were two periods of Precambrian deformation: (i) between the middle Proterozoic platformal rocks (Green Head = Gamble Brook = George River Groups) and the late Proterozoic volcano-sedimentary rocks (Coldbrook = Jeffers = Folly River = Fourchu Groups); and (ii) between the late Proterozoic volcano-sedimentary rocks and the Cambrian rocks. Other workers (e.g. Helmstaedt and Tella, 1973; Keppie, 1982) believed that the George River and Fourchu Groups were concordant, and that some of the deformational fabrics were produced just prior to deposition of the overlying Cambrian rocks while others are attributed to Paleozoic deformation. A third alternative was presented by Nance (1982) who believed that all the structures observed in the Green Head Group are of Paleozoic age. Several types of data bearing on these problems are available in selected areas.

In the Caledonian Highlands of southern New Brunswick, the Brookville Gneiss records upper amphibolite facies metamorphism accompanied by polyphase deformation: complex small interference patterns formed by roughly coplanar axial planes but variably oriented fold axes in narrow linear zones within more homogeneous gneiss with planar or gently curved to pinch-and-swell gneissosity (Currie *et al.*, 1981). This deformation is bracketed between the ages of the youngest detrital zircon separated from paragneiss (641 ± 3 Ma: 1% discordant Pb/Pb age: Bevier *et al.*, 1990) and metamorphic titanite (564 ± 6 Ma: nearly concordant Pb/Pb age: Bevier *et al.*, 1990).

In the central Cobequid Highlands, the Great Village River Gneiss is composed of polydeformed hornblende-bearing orthogneisses, biotite-garnet psammitic paragneisses and quartz-plagioclase layered amphibolite dykes. A 734 ± 3 Ma U-Pb zircon age from hornblende orthogneisses (Economy River gneiss) provides a minimum age for the Gamble Brook Formation xenoliths of which occur in the orthogneiss and a maximum age for the polyphase deformation (Doig *et al.*, 1991). Syntectonic granite gneisses and amphibilites intrusive into the Great Village Gneiss have yielded 600-580 Ma U-Pb ages (Doig *et al.*, 1991), and 625-640 Ma Rb-Sr isochron ages (Gaudette *et al.*, 1984). Intense mylonitic fabrics in the Gamble Brook Formation (= Middle Proterozoic Green Head Group) are truncated by a mafic dyke intruded along the contact with the late Proterozoic Folly River Formation (Nance and Murphy, 1990). C-S fabrics, gently E-SE plunging stretching lineations defined by dimensionally oriented hornblende, quartz and quartz-feldspar augen, and fold asymmetry in a moderately-steeply SE-dipping mylonitic fabric indicate an oblique sinistral sense of shear (Nance and Murphy, 1990). Xenoliths of mylonitic Gamble Brook quartzite have been recorded in the post-tectonic ca. 609 Ma old Debert River pluton (Doig *et al.*, 1991). A mafic dyke that truncates fabrics in the Gamble Brook Formation is petrographically and chemically similar to the mafic lavas in the Folly River Formation and is inferred to be a feeder dyke. This suggests that the contact between the Gamble Brook and Folly River formations is an unconformity representing deformation. The late Proterozoic rocks of the Jeffers Group and Folly River Formation were deformed by thrusts and isoclinal folds with kinematics indicating dextral transpression (Nance and Murphy, 1990). This deformation took place prior to the intrusion of the Debert River pluton at ca. 609 Ma (U-Pb zircon age; Doig *et al.*, 1991) and

596 ± 70 Ma (Rb- Sr whole-rocks isochron; Donohoe *et al.* 1986). Thus, the deformational events in the Cobequid Highlands are constrained between ca. 615 and 580 Ma.

A similar relationship has been recorded in the northern Antigonish Highlands where polyphase fabrics in the Georgeville Group are truncated by the Greendale Complex (Murphy *et al.*, 1982 and in press). This deformation is constrained between the age of the youngest detrital zircon in the Georgeville Group (ca. 613 Ma: Keppie and Krogh, 1990) and the oldest $^{40}Ar/^{39}Ar$ hornblende plateau age in the Greendale Complex (620 ± 5 Ma: Keppie *et al.*, 1990). This polyphase deformation dies out southwards and the Georgeville Group in the southern Antigonish Highlands is only mildly deformed (Murphy *et al.*, 1991).

In Cape Breton Island the type George River and Fourchu Groups are not in contact. However, Helmstaedt and Tella (1973) recorded volcanic fragments resembling the Fourchu Group in a limestone breccia of the George River Group. They concluded that the Fourchu volcanism began during a late stage of the George River sedimentation. In the Craignish Gills and North Mountain, volcanic rocks have been recorded within the George River Group and may represent a volcanic episode distinct from the Fourchu Group (Parsons, 1964; Milligan, 1970). The oldest volcanic rocks in the Fourchu Group are 676 ± 1 Ma (U-Pb zircon age: Barr *et al.*, 1990) considerably younger than the inferred Middle Proterozoic age of the George River Group (based upon a litho-stratigraphic correlation with the fossiliferous Green Head Group).

The oldest rocks exposed in the Avalon Zone in Newfoundland are the oceanic, tholeiitic, mafic volcanics of the Burin Group (Strong *et al.*, 1978). A gabbroic sill in this unit has yielded a U-Pb age of 760 ± 2 Ma (Krogh, 1982). These rocks are inferred to have formed in an aborted rift (Strong *et al.*, 1978). A series of ESE-WNW, E-W to NE-SW, dextral wrench faults, which only affect the Burin Group but not the younger Marystown Group, have been tentatively correlated with this rifting event (Strong *et al.*, 1978).

Along the southern coast of Newfoundland (Burgeo Terrane), amphibolite facies gneisses (Cinq Cerf gneiss) display a penetrative foliation associated with isoclinal folds in gneissic and migmatitic banding (Blackwood, 1985). This deformation and metamorphism are bracketed by U-Pb data from migmatitic gneiss which yielded a concordant zircon point at 663 ± 3 Ma interpreted as the time of intrusion and a concordant titanite age of 579 ± 10 Ma that is inferred to closely follow peak metamorphism (Dunning and O'Brien, 1989). An upper limit on the high grade tectonothermal event is given by (1) post-tectonic granites that yielded intrusive ages of 563 ± 4 and 499 +3/-2 Ma (Dunning and O'Brien, 1989), and (2) unconformably(?) overlying low grade metasedimentary and metavolcanic rocks yielding magmatic zircon of 544 ± 5 Ma age and detrital zircon of c. 593 Ma age (Dunning and O'Brien, 1989; O'Brien *et al.*, 1989).

In central Newfoundland, late Proterozoic plutons are exposed in two tectonic windows in the surrounding rocks of the Exploits subzone (Evans *et al.*, 1990). These plutons yielded U-Pb zircon crystallization ages of 563 ± 2 and 565 +4/-3 Ma and a concordant U-Pb monazite age of 545 ± 3 Ma interpreted as a metamorphic age. The nearest plutons of similar age lie within the Burgeo Terrane.

Stratigraphic and sedimentological data place an upper age limit on some of the deformation. Cambrian rocks rest unconformably upon polydeformed Late Precambrian Georgeville Group in the northern Antigonish Highlands (Murphy *et al.*, 1982 and 1991). In southern New Brunswick, Early Cambrian rocks rest disconformably upon the Coldbrook Group. Elsewhere critical contacts are not exposed, however clasts deformed prior to incorporation in Cambrian rocks have been recorded in several places. Thus, Ruitenberg *et al.* (1979) and McCutcheon (1981) reported deformed clasts of both the Green Head and Coldbrook Groups in Lower Cambrian conglomerates of New Brunswick. Similarly, deformed Georgeville clasts occur in the Lower Cambrian conglomerates of the Antigonish Highlands (Murphy *et al.*, 1982 and 1991). In the Boisdale Hills of southeastern Cape Breton Island, George River clasts containing post-tectonic chiastolite crystals enclosing two fabrics have been found in Middle Cambrian conglomerates (Helmstaedt and Tella, 1973). These chiastolite crystals were related to contact metamorphism of the Boisdale Hills Pluton which has yielded a U-Pb age of 564 +3/-2 Ma (Barr *et al.*, 1990), a Rb-Sr whole rock isochron of 563 ± 31 Ma (Cormier, 1972; recalculated by Keppie and Smith, 1978), and $^{40}Ar/^{39}Ar$ plateau ages on hornblende ranging between 527 ± 6 and 535 ± 2 Ma (Keppie *et al.*, 1990). In southeastern Cape Breton Island, pebbles derived from the Fourchu Group and displaying an inherited fabric occur in the Early Cambrian Morrison River Formation (Keppie, 1982; Smith, 1978). In the Newfoundland Avalon Composite Terrane, abundant detrital mica occurs in the Upper Cambrian and Ordovician rocks and detrital muscovite and garnet and metamorphic rock fragments have been recovered from the upper Late Precambrian (Papezik, 1973; Jenness, 1963; Blackwood and Kennedy, 1975; Blackwood and O'Driscoll, 1976 and Kennedy, 1976).

Some of the deformational events affecting the Avalon Zone in Nova Scotia and New Brunswick appear be reflected stratigraphic record in the Avalon Zone in Newfoundland. Thus, the Conception = Connecting Point Groups have been interpreted as flysch (King, 1980) and if so are syntectonic. Both Riphean and Vendian ages have been inferred for these groups based upon correlations of the tillite (near the base) and Ediacaran fauna (near the top)(Anderson, 1972, 1978; Bruekner *et al.*, 1977; Williams and King, 1979). A lower limit on flysch deposition is given by: (i) a rhyolite porphyry pebble in the Rock Harbour Group (= Musgravetown Group) that has yielded a 623 ± 2 Ma U-Pb zircon age (Krogh *et al.*, 1988): this pebble could have been derived from the Harbour Main Group, which lies stratigraphically beneath the Conception Group, and is also dated at 623 ± 2 Ma and 606 ± 4 Ma by U-Pb (Krogh *et al.*, 1988); (ii) the tillite unit of the Conception Group contains granite pebbles similar to the Holyrood granite which is 621 ± 2 Ma old (U-Pb, Krogh *et al.*, 1988); and (iii) the Love Cove Group, lying stratigraphy beneath the Connecting Point Group, has been dated by U-Pb on zircon at 608 ± 25 Ma and 590 ± 30 Ma (Dallmeyer, 1980; Dallmeyer *et al.*, 1983), which is similar to the 580 ± 20 Ma U-Pb age for the (?)subvolcanic Swift Current granite (op. cit.). An upper limit on the age of the Conception = Connecting Point Groups is provided by (1) the 565 ± 3 Ma U-Pb zircon age from a tuff in the Mistaken Point Formation (the uppermost unit of the Conception Group; Benus, 1988); (2) the 608+20/-8 Ma (U-Pb on zircons, Krogh *et al.*, 1988) on an ignimbrite of the Marystown Group which lies (?)disconformably beneath the Musgravetown Group. Thus, if the flysch is syntectonic then its deposition between ca. 625-560 Ma also dates the time of deformation.

The 8-9 km thick St. John, Signal Hill, Hodgewater and Musgravetown Groups overlying the Conception - Connecting Point Groups have been interpreted as molasse (King, 1980). While they contain deformed and metamorphosed detritus and thus post-date some deformation, molassic units are generally synchronous with deformation. They were probably deposited between c.560 Ma and the base of the Cambrian at c.530-545 Ma. The base of the Cambrian is placed at about 570 Ma by Palmer (1983), although Odin (1982) places it at 530 ± 10 Ma. A recent assessment of the base of the Cambrian places it between 530 and 545 Ma (Keppie *et al.*, 1990). This agrees with the oldest and most tightly constrained Rb-Sr whole rock isochron of 574 ± 11 Ma on the Shunacadie Pluton.

Collectively, these stratigraphic/sedimentological data and intrusive relations provide evidence for deformation intermittently between ca. 625 Ma and the base of the Cambrian. These deformational events are not pervasive throughout the Avalon Zone, and appear to be sporadically developed, in some cases associated with discrete shear zones. Much of this deformation is contemporaneous with extrusion of the late Proterozoic Fourchu and Coldbrook groups. The geochemistry of these volcanic units indicates that they were extruded in a volcanic arc setting (Dostal *et al.*, 1990; Dostal and McCutcheon, 1990). The Fourchu Group ranges in ages from 676 ± 1 to 574 ± 1 Ma (U-Pb zircon ages: Barr *et al.*, 1990), comparable to that of the Coldbrook Group: 635-600 Ma (U-Pb zircon ages: Bevier and Barr, 1990). In such a generally convergent tectonic environment, deformation probably took place throughout the period of subduction.

There is evidence that deformation continued into the Cambrian and Ordovician. This is expressed as local unconformities and the nonconformity between the basal Cambrian and the Holywood granite (representing 6-20 km of uplift and erosion: Hughes, 1971; Strong and Minitiades, 1976) of fault-bounded horsts such as the Holyrood and Isthmus (McCartney, 1969). Cambrian-Ordovician movements may be recorded in the disconformities within the Cambrian- Ordovician rocks. In Newfoundland, Hutchinson (1962) recorded the following diastems: (i) between the Early and Middle Cambrian marked by a Middle Cambrian manganese horizon in the north, however beds missing in the north are present in the southwest; (ii) within the late Middle Cambrian appearing as a phosphatic clast conglomerate at the base of the Elliot Cove Formation (King, 1982) and (iii) in the Early Ordovician at the base of the Arenigian Wabana Group (Ranger, 1979). In southern Cape Breton Island, although contacts are not exposed, a disconformity is believed to occur between the Middle and Upper Cambrian rocks (Hutchinson, 1952). Three paleontological zones are missing in the Mira area, whereas six are absent in the Boisdale Hills. In the Antigonish Highlands, there appears to be a disconformity between the Lower Cambrian and (?)Lower Ordovician rocks (Murphy *et al.*, 1982 and 1991). However, the structure here is complex (Keppie and Murphy, 1988) and the outcrop is rather limited so that one cannot be certain whether the break is stratigraphic or structural. In southern New Brunswick, the Cambrian-Ordovician is almost complete with only one faunal zone missing at the top of the Middle Cambrian (Hayes and Howell, 1937). In the Antigonish Highlands, the relative motion on NE-SW trending faults bounding the pull-apart basin was inferred to have been dextral during the Early Cambrian (Keppie and Murphy, 1988).

HUMBER ZONE

Structures of Late Hadrynian-Cambrian age in the Humber zone are mainly normal faults associated with the formation of Iapetus. These faults form two groups: (i) those bordering Iapetus and (ii) those associated with the failed arms or aulacogens of rrr triple junctions. The normal faults bordering Iapetus in Quebec trend NNE-SSW and E-W and have downthrows on their southeastern side. Times of active faulting are indicated by the fact that the faults form the northwestern limit to Late Precambrian and Lower Cambrian strata (St. Julien and Hubert, 1975). Subsequent late Lower Cambrian to Lower Ordovician rocks (and Middle Ordovician at Quebec City) transgressed progressively northwestwards across the Grenvillian basement apparently without any major breaks in sedimentation. In Newfoundland, these early normal faults are obscured by subsequent nappe-type deformation, however, the major transgressive event appears to have taken place during late Lower Cambrian times (Rogers, 1967; Stevens, 1976; Williams and Hiscott, 1987). Sedimentation was essentially continuous throughout the Cambrian with only minor breaks in the Middle Cambrian at Port-au-Port and Hare Bay (Stevens, 1965; Kay, 1967). In Newfoundland, syn-rift, NNE-SSW, tholeiitic, mafic, dyke feeders to basalts locally swing westwards at the latitude of the Baie Verte Flexure (Hibbard, 1983a). They yielded K-Ar ages in the range of 766 to 876 Ma (Pringle *et al.*, 1971). Subsequently, Stukas and Reynolds (1974a) produced ^{40}Ar/^{39}Ar ages of 615 ± 10 Ma and suggested that the older ages were the result of excess argon. Cooling ages from the host Grenvillian gneisses of the Indian Head Range and the Long Range of 790 - 860 Ma (^{40}Ar/^{39}Ar, Dallmeyer, 1978) and 955 - 970 Ma (K-Ar mica, Lowdon *et al.*, 1963) respectively, suggest that the 615 ± 10 Ma dates probably represent the age of dyke intrusion. This is consistent with the age of a granite at Hughes Lake correlated with the bimodal rift volcanism which has yielded a U-Pb age of 602 ± 10 Ma (Williams *et al.*, 1985).

Failed arms associated with Iapetus occur at Ottawa, Saguenay and Cape Farewell and are marked by faults and c.550 Ma old alkaline intrusions (Doig, 1970). The E-W to ESE-WNW, Ottawa-Bonnachere graben and associated dyke swarm may be traced westwards 300 km to North Bay and contains Upper Cambrian and Ordovician sediments which thin from 1 km near Montreal, through 0.5 km near Ottawa, to zero about 150 km farther west (Kay, 1942).

INTERPRETATION

The structural data for the Humber Zone suggest that Iapetus originated by initial lithospheric stretching (McKenzie, 1978) followed by lithospheric dyking (Royden *et al.*, 1980). The rifting phase occurred in late Proterozoic times with the tholeiitic mafic dykes appearing slightly before the birth of Iapetus oceanic crust. Subsequent lithospheric cooling led to subsidence of the margin and sedimentation during the early Cambrian causing flexuring of the lithosphere and a peripheral bulge on the craton. The NNE-SSW to E-W trends of the dyke swarms and normal faults are parallel to the margins of the Grenvillian craton. This indicates that the rifting was oblique or transtensional, with the relative plate motion oriented somewhere between E-W through N-S to NNE-SSW.

Late Proterozoic structures in the Avalon Zone vary from disconformities to polyphase fabrics. In Newfoundland and parts of southern New Brunswick, the major structure produced during this late Proterozoic event in the Avalon Zone involved block faulting, tilting and warping. In Nova Scotia, the structure appears to consist of a series of horsts separated by grabens, such as the Antigonish-Cobequid Highlands graben (Keppie, 1982). The Avalon Zone has generally been interpreted as a rifted magmatic arc complex produced by oblique subduction (Keppie, 1982; Keppie *et al.*, 1991; Keppie and Dostal, 1991). In this context, deformation would be limited to an area adjacent to intra-arc shear zones, which would explain its sporadic development. Bends in intra-arc transcurrent faults would also explain the synchroneity of rifting, transcurrent faulting and compressive deformation. Thus, deposition of flysch in the southern Antigonish Highlands appears to be synchronous with deformation and intrusion in the northern Antigonish Highlands. On a regional scale deposition of the flysch in the Antigonish and Cobequid Highlands and in eastern Newfoundland appears to be synchronous with the deformation in other areas (Keppie *et al.*, 1991). Geochronological data suggests that the magmatic arc lasted from c.680-560 Ma. The polarity of subduction was towards the northwest beneath southern Cape Breton Island and southern New Brunswick (Dostal *et al.*, 1990; Dostal and McCutcheon, 1990; Keppie and Dostal, 1991). The geochemistry of the Late Proterozoic igneous rocks indicates that the magmatic arc was built upon continental crust. The continental basement is only exposed in the Caledonian and Cobequid Highlands and in northern Cape Breton Island, where it consists of a middle Proterozoic miogeocline possibly resting unconformably upon a middle Proterozoic gneissic basement in Cape Breton Highlands. That transpression following the cessation of subduction probably continued into Cambrian times is indicated by the presence of localized, narrow zones of bimodal, tholeiitic or alkalic, continental rift volcanic rocks of Early - (?)Late Cambrian age and the presence of diastems (Keppie, 1982; Murphy *et al.*, 1985, Greenough and Papezik, 1985; Greenough *et al.*, 1985).

Faunal provinciality and paleomagnetic data indicate that the Avalon Zone had affinities with Gondwana and Armorica during Late Proterozoic-Cambrian times (Johnson and Van der Voo, 1985; 1986). In most of these regions the Late Proterozoic orogens are built upon Archean and Early Proterozoic rocks. Only in South America are Late Proterozoic orogens built upon a Middle-Late Proterozoic (Grenvillian) basement (Keppie and Dallmeyer, 1989), which may, therefore, represent the provenance of the Avalon Zone. The synchroneity of Late Proterozoic subduction beneath the Avalon Zone and the opening of Iapetus implies that the Avalon was not within the Iapetus Ocean. The good state of preservation of the Avalonian magmatic arc sequences contrasts with their general elimination in continent-continent collision orogens where they are usually eroded away because they lie on the upper plate. This suggests that the termination of subduction beneath the Avalon Zone was not due to continental collision. Instead, it may have ended as a result of the relocation of subduction zones into Iapetus as it evolved from a passive to an active margin ocean in the Late Cambrian.

PALEOZOIC STRUCTURES: APPALACHIAN OROGENY

Structures of Paleozoic age are responsible for most of the present geometry of the Canadian Appalachians (Fig. 1). In the Humber, Dunnage and Gander Zones, the Cambrian Period appears to represent a relatively quiescent interval between the end of Proterozoic deformation and the onset of Paleozoic deformation, except for the presence of ca. 513 Ma old calcalkaline volcanic rocks in central Newfoundland (Evans *et al.*, 1990). The nature of the structures and constraints on their age are illustrated in five transects across the orogen (Figs. 4-8). It is clear that deformation has been relatively continuous from Ordovician to Permian. It follows from this observation that the Appalachians were produced by just one orogeny, the Appalachian Orogeny. Subdivision of the Appalachian Orogeny into deformational episodes is based upon spatial or temporal breaks in the deformation and reversals of vergence. On two of the transects (Figs. 5 & 7), Ordovician deformation appears to be absent close to the axis of the orogen across which the Ordovician vergence changes, i.e. across a line from Green Bay to Cape Ray in Newfoundland and from Port Daniel to Edmundston in New Brunswick (Fig. 1). Along the northwestern side of the orogen, a temporal break is clearly present between the end of Ordovician deformation and the onset of Devonian deformation. On the other hand, along the Gander and southeastern Dunnage Zones deformation appears to be essentially continuous, however the vergence changes: south to southeastwards in the Early Paleozoic and northwestwards to westwards in the Late Paleozoic (Figs. 5-7). This provides the basis for the subdivision of the Appalachian Orogeny into several redefined deformational episodes (Figs. 9-11):

Taconian: Ordovician Deformation Northwest of the Axis of Vergence Reversal;

Penobscotian: Early Ordovician deformation between the Exploits subzone and the Gander Zone;

Beothukan: Late Ordovician, Silurian and Early Devonian deformation between the Notre Dame and Exploits subzones;

Ganderian: Silurian deformation of the eastern Dunnage, Gander and Avalon zones;

Acadian and Alleghanian: Late Paleozoic deformation with a northwest-west vergence that affects the entire orogen.

Taconian Episode of Deformation: Early-Late Ordovician and Northwesterly Vergent in the Humber and Western Dunnage Zones (Notre Dame/Dashwood Subzones)

The Taconian episode of deformation is defined by the structures of Ordovician age found along the northwestern half of the Appalachian Orogen in the Humber and northwestern parts of the Dunnage Zones (Figs. 4-9). The major structure consists essentially of a stack of allochthonous slices which from bottom to top are: (1) cratonic basement overlain by miogeoclinal and flyschoid rocks; (2) continental rise prism; (3) ocean floor rocks; and (4) the uppermost slice containing ophiolite in the west and oceanic/cratonic volcanic arc or volcanic island in the east (Figs. 12-16). The slices are usually separated by melange which is commonly ophiolitic. The oldest structures occur at the base of the ultramafic rocks, and all appear to be Tremadocian: no diachronism along the orogen was detected. The age of the earliest structures becomes progressively younger downwards through the stacked allochthonous slices, and some later superimposed deformation may be detected at higher levels. The Taconian deformational episode appears to have ended diachronously along the Canadian Appalachians: Ashgillian in Quebec and Caradocian in Newfoundland (Fig. 17), although continued deformation at a deeper structural level in Newfoundland cannot be discounted. The nature of these structures and their age constraints will be described below followed up by a discussion of their genesis.

HUMBER ZONE

Ophiolite

Ophiolitic rocks, in the sense of Coleman (1977), occur in the uppermost structural slice of several allochthons thrust over the miogeoclinal rocks of the Humber Zone. While some of these ophiolites have been thrust far out over the continental shelf (e.g., Bay of Islands), others lie above the continental rise prism (e.g. Thetford, Mings Bight). Structures in some of these rocks have received detailed study. The ophiolites represent fragments of oceanic lithosphere comprising from base to top: ultramafic tectonites, layered ultramafic and gabbroic rocks, in part cumulates; layered to massive gabbros; sheeted diabase dykes and altered mafic pillow lavas; capped by a thin sequence of sedimentary rocks. The ultramafic rocks and gabbros display a variety of structural features including compositional banding in the peridotites; magmatic accumulation planes and magmatic lineation in the gabbros and a foliation ± mineral lineation, which are commonly deformed by folds ± associated foliation and mineral lineation in the basal peridotite tectonites (Girardeau and Nicolas, 1981; Mercier, 1976; Christensen and Salisbury, 1979; Girardeau, 1982; MacGregor and Basu, 1979; Beaudin, 1980; Clague *et al.*, 1981; Laurent, 1975, 1977). The earliest foliation is generally parallel to the lherzolite-harzburgite compositional layering and is defined by the preferred orientation of tabular minerals such as orthopyroxene and chromite, whereas the associated lineation is marked by aligned long axes of spinel, pyroxene and plagioclase. Olivine crystallographic axes display strong preferred crystallographic orientations with: (i) in the porphyroclastic peridotites, [010] perpendicular to the foliation and [100] parallel to the lineation, or (ii) in the lower mylonitic peridotites, [001] perpendicular to the foliation and [100] parallel to the lineation. The former indicates rotational deformation with a dominant (010) [100], moderate to high temperature slip system at low stress, whereas, the latter suggests rotational deformation with a dominant (100) [001], relatively low temperature, high stress slip system (Carter and Ave Lallemant, 1970). Recrystallization of olivine, probably by subgrain rotation, in the porphyroclastic peridotite produced neoclasts averaging 90-110 μm with some between 175-275 μm. A bimodal size distribution of olivine and

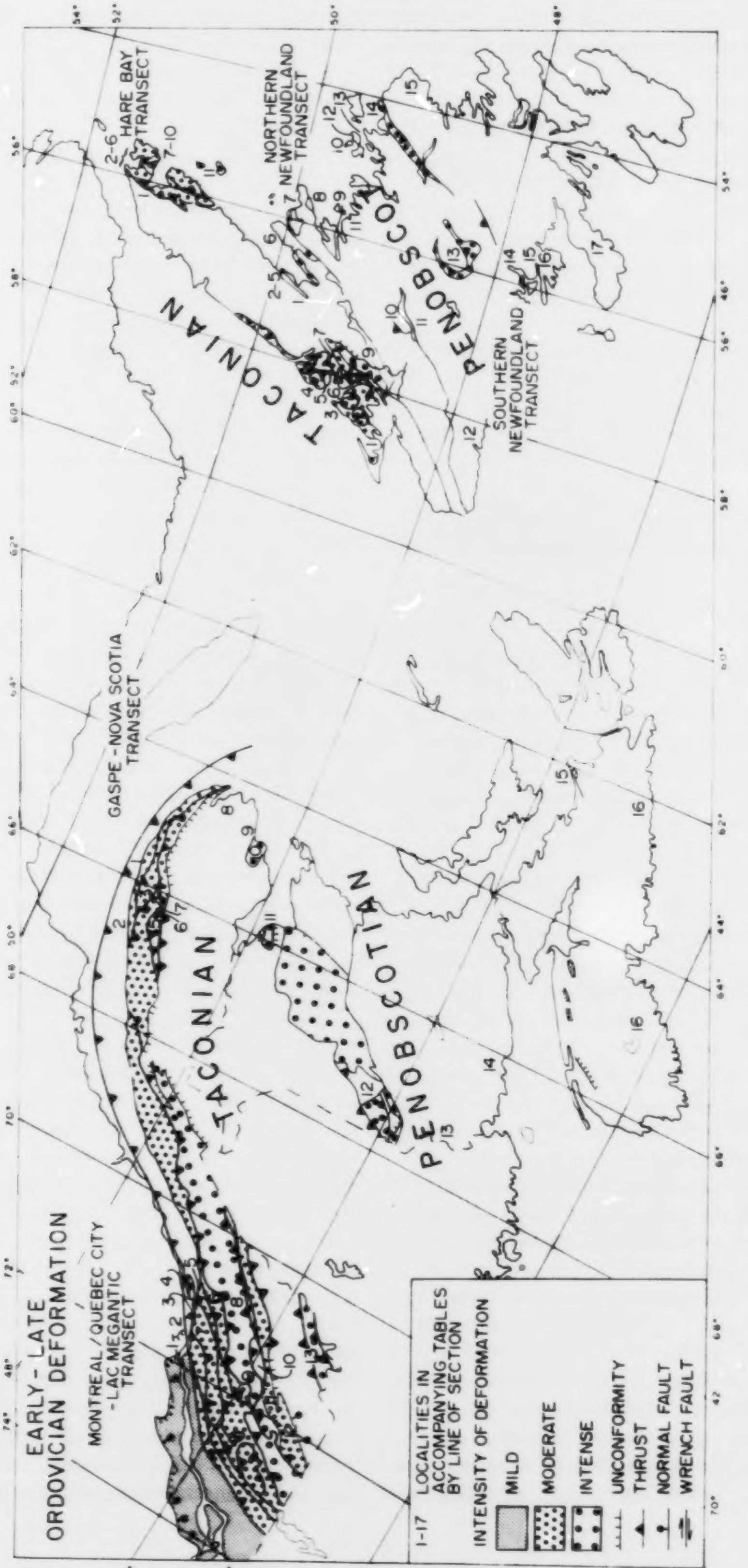

EARLY - LATE ORDOVICIAN DEFORMATION
MONTREAL / QUEBEC CITY -LAC MEGANTIC TRANSECT
GASPE - NOVA SCOTIA TRANSECT
HARE BAY TRANSECT
NORTHERN NEWFOUNDLAND TRANSECT
SOUTHERN NEWFOUNDLAND TRANSECT
TACONIAN
PENOBSCOTIAN
1-17 LOCALITIES IN ACCOMPANYING TABLES BY LINE OF SECTION
INTENSITY OF DEFORMATION
MILD
MODERATE
INTENSE
UNCONFORMITY
THRUST
NORMAL FAULT
WRENCH FAULT

Figure 9. Map of Early-Late Ordovician structures in the Canadians Appalachians.

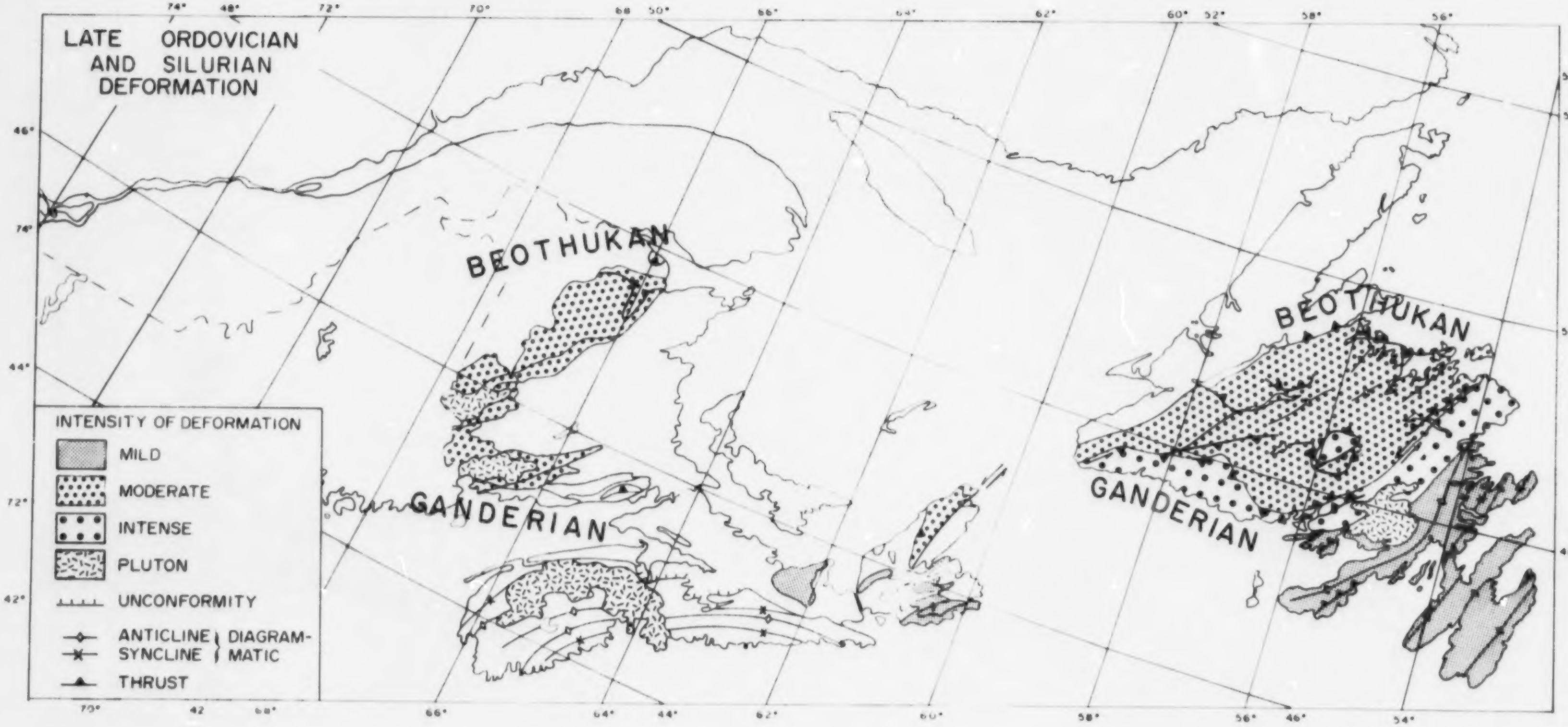

LATE ORDOVICIAN AND SILURIAN DEFORMATION
BEOTHUKAN
GANDERIAN
BEOTHUKAN
GANDERIAN
INTENSITY OF DEFORMATION
MILD
MODERATE
INTENSE
PLUTON
UNCONFORMITY
ANTICLINE } DIAGRAMMATIC
SYNCLINE
THRUST
74°
48°
72°
70°
68
50°
66°
64°
62°
60°
52°
58°
56°
54°
52°
50°
48°
46°
74°
44°
72°
42°
70°
42
68°
66°
64°
44°
62°
60°
58°
56°
46°
54°

Figure 10. Map of Late Ordovician and Silurian structures in the Canadian Appalachians.

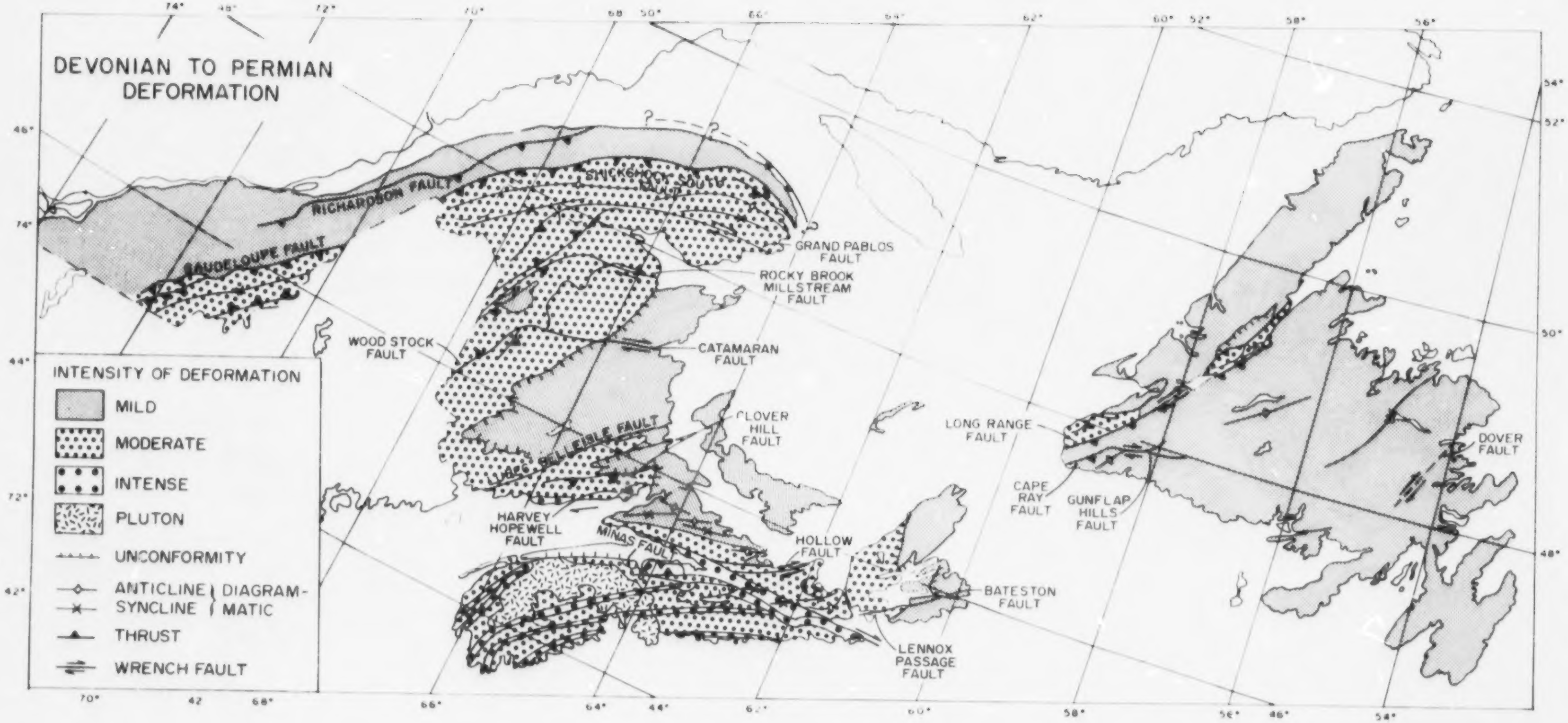

DEVONIAN TO PERMIAN DEFORMATION
INTENSITY OF DEFORMATION
MILD
MODERATE
INTENSE
PLUTON
UNCONFORMITY
ANTICLINE } DIAGRAMMATIC
SYNCLINE
THRUST
WRENCH FAULT
RICHARDSON FAULT
GAUDELOUPE FAULT
SHICKSHOCK SOUTH FAULT
GRAND PABLOS FAULT
ROCKY BROOK MILLSTREAM FAULT
WOOD STOCK FAULT
CATAMARAN FAULT
CLOVER HILL FAULT
BELLEISLE FAULT
HARVEY HOPEWELL FAULT
MINAS FAULT
HOLLOW FAULT
BATESTON FAULT
LENNOX PASSAGE FAULT
LONG RANGE FAULT
CAPE RAY FAULT
GUNFLAP HILLS FAULT
DOVER FAULT
74°
48°
72°
70°
68
50°
66°
64°
62°
60°
52°
58°
56°
54°
46°
44°
42°
42
54°
50°
48°
46°
44°

Figure 11. Map of Devonian to Permian structures in the Canadian Appalachians.

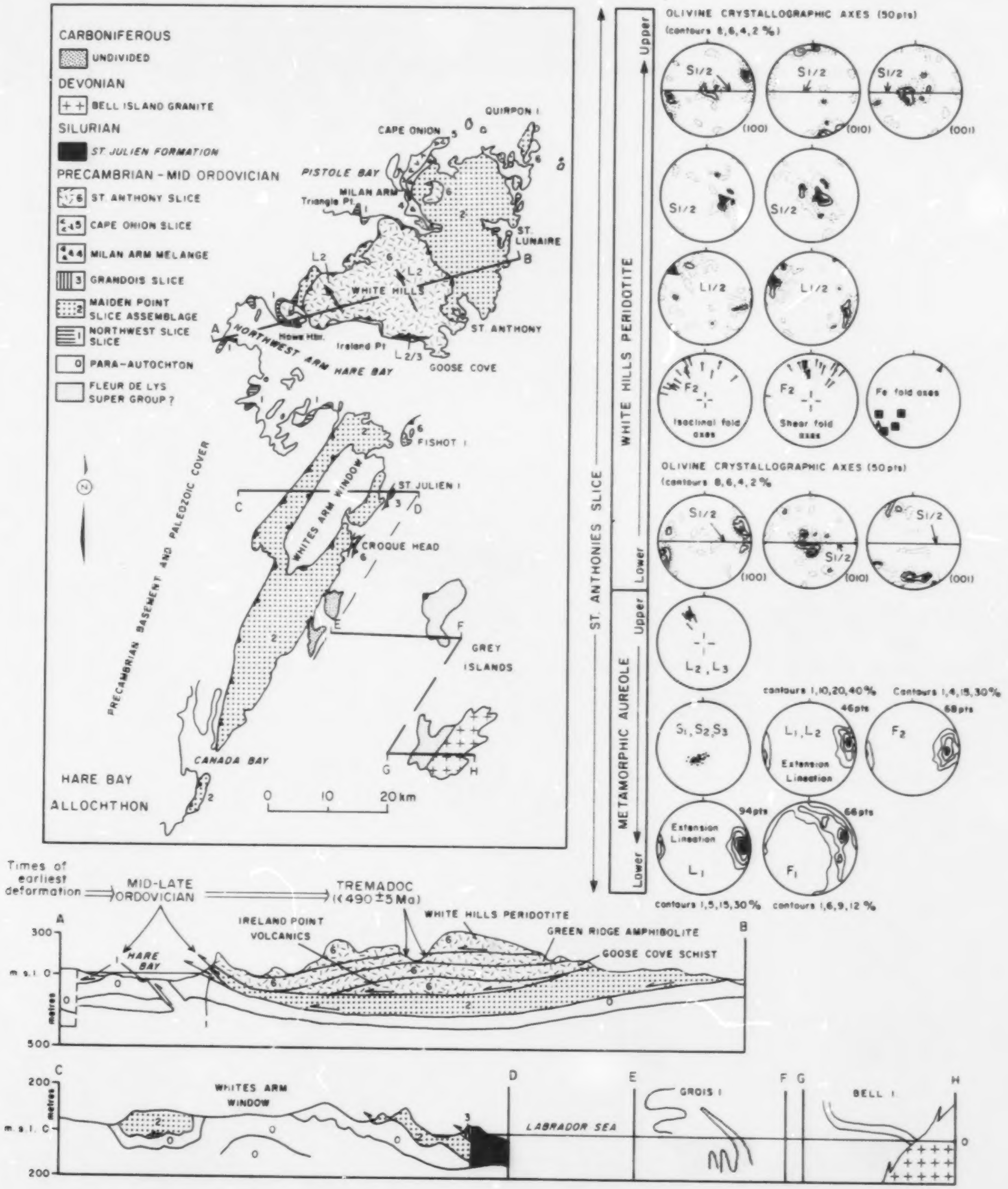

CARBONIFEROUS
UNDIVIDED
DEVONIAN
BELL ISLAND GRANITE
SILURIAN
ST JULIEN FORMATION
PRECAMBRIAN - MID ORDOVICIAN
6 ST. ANTHONY SLICE
5 CAPE ONION SLICE
4 MILAN ARM MELANGE
3 GRANDOIS SLICE
2 MAIDEN POINT SLICE ASSEMBLAGE
1 NORTHWEST SLICE SLICE
0 PARA-AUTOCHTON
FLEUR DE LYS SUPER GROUP?
QUIRPON I
CAPE ONION
PISTOLE BAY
MILAN ARM
Triangle Pt.
ST. LUNAIRE
WHITE HILLS
ST ANTHONY
Howe Hbr.
Ireland Pt
GOOSE COVE
NORTHWEST ARM HARE BAY
FISHOT I
ST JULIEN I
WHITES ARM WINDOW
CROQUE HEAD
GREY ISLANDS
CANADA BAY
PRECAMBRIAN BASEMENT AND PALEOZOIC COVER
HARE BAY ALLOCHTHON
0 10 20km
N
ST. ANTHONIES SLICE
WHITE HILLS PERIDOTITE
METAMORPHIC AUREOLE
Upper
Lower
OLIVINE CRYSTALLOGRAPHIC AXES (50pts)
(contours 8,6,4,2 %)
S1/2
(100)
(010)
(001)
L1/2
F2
Isoclinal fold axes
Shear fold axes
F2 fold axes
L2, L3
contours 1,10,20,40%
Contours 1,4,15,30%
46pts
68pts
S1, S2, S3
L1, L2
Extension Lineation
94pts
66pts
L1
F1
contours 1,5,15,30%
contours 1,6,9,12 %
Times of earliest deformation
MID-LATE ORDOVICIAN
TREMADOC (490 ± 5 Ma)
A
B
WHITE HILLS PERIDOTITE
IRELAND POINT VOLCANICS
GREEN RIDGE AMPHIBOLITE
GOOSE COVE SCHIST
HARE BAY
300
m.s.l. 0
500
metres
C
D
E
F
G
H
WHITES ARM WINDOW
GROIS I
BELL I
LABRADOR SEA
200
m.s.l. 0
200
metres

Figure 12. Map and section of the Hare Bay Allochthon and Grey Islands to show the main structural slices (after Williams, 1975 & Williams & Smyth, 1973; Kennedy *et al*., 1973). Also shown are the fabric elements in the White Hills peridotite (after Girardeau, 1982) and in the underlying metamorphic aureole at Starks Bight (after Craw, 1983), both parts of the highest St. Anthony slice.

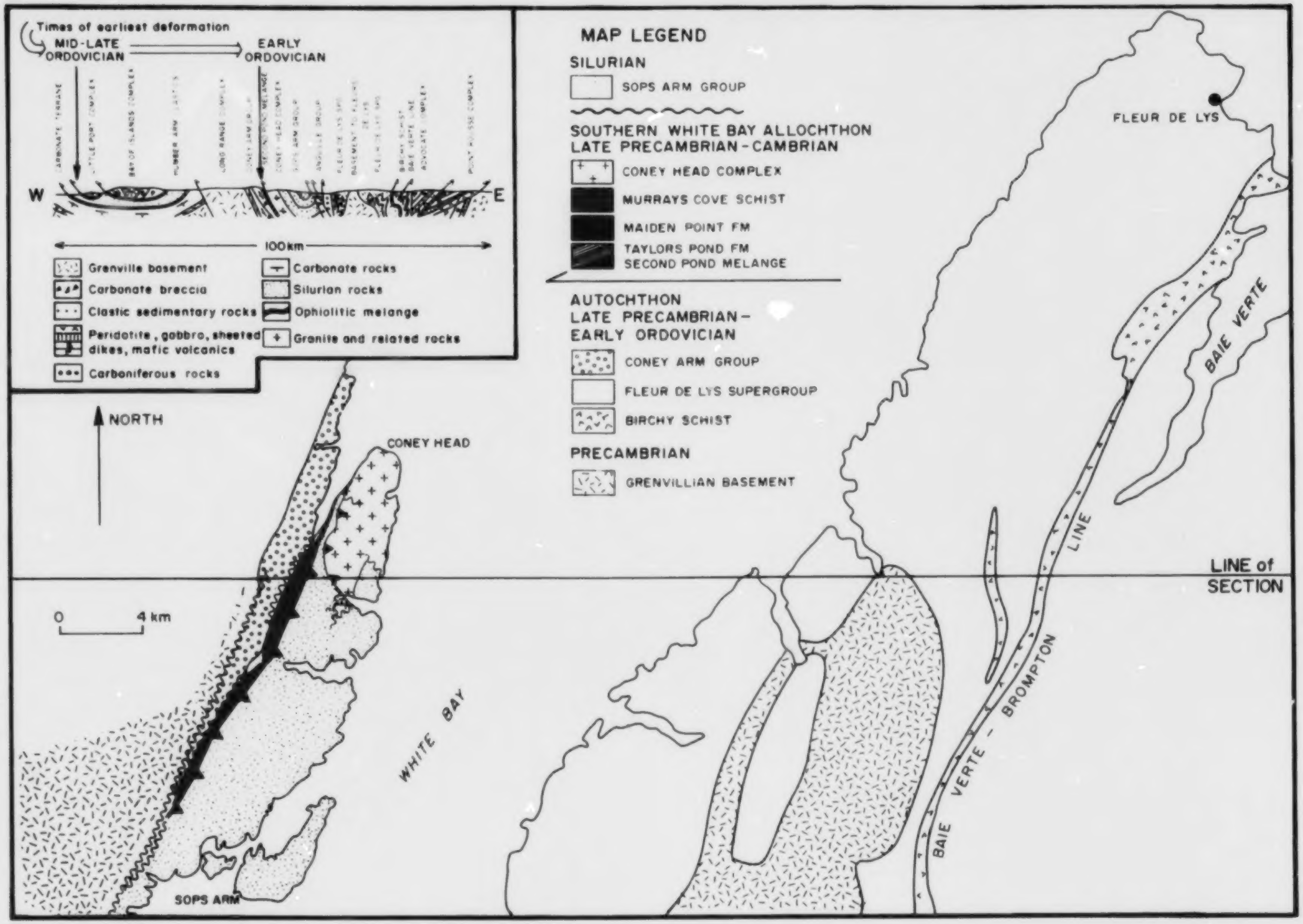

Times of earliest deformation
MID-LATE ORDOVICIAN
EARLY ORDOVICIAN
CARBONATE TERRANE
LITTLE PORT COMPLEX
BAY OF ISLANDS COMPLEX
HUMBER ARM CLASTICS
LONG RANGE COMPLEX
CONEY ARM GROUP
SECOND POND MELANGE
CONEY HEAD COMPLEX
SOPS ARM GROUP
ANGUILLE GROUP
FLEUR DE LYS SPG
BASEMENT TO FLEUR DE LYS
FLEUR DE LYS SPG
BIRCHY SCHIST
BAIE VERTE LINE
ADVOCATE COMPLEX
POINT ROUSSE COMPLEX
W
E
100km
Grenville basement
Carbonate breccia
Clastic sedimentary rocks
Peridotite, gabbro, sheeted dikes, mafic volcanics
Carboniferous rocks
Carbonate rocks
Silurian rocks
Ophiolitic melange
Granite and related rocks
MAP LEGEND
SILURIAN
SOPS ARM GROUP
SOUTHERN WHITE BAY ALLOCHTHON
LATE PRECAMBRIAN - CAMBRIAN
CONEY HEAD COMPLEX
MURRAYS COVE SCHIST
MAIDEN POINT FM
TAYLORS POND FM
SECOND POND MELANGE
AUTOCHTHON
LATE PRECAMBRIAN - EARLY ORDOVICIAN
CONEY ARM GROUP
FLEUR DE LYS SUPERGROUP
BIRCHY SCHIST
PRECAMBRIAN
GRENVILLIAN BASEMENT
NORTH
0
4 km
CONEY HEAD
WHITE BAY
SOPS ARM
FLEUR DE LYS
BAIE VERTE
BAIE VERTE - BROMPTON LINE
LINE of SECTION

Figure 13. Map and section of the Southern White Bay Allochthon and western Fleur de Lys Supergroup (after Smyth & Schillereff, 1982, Williams, 1977).

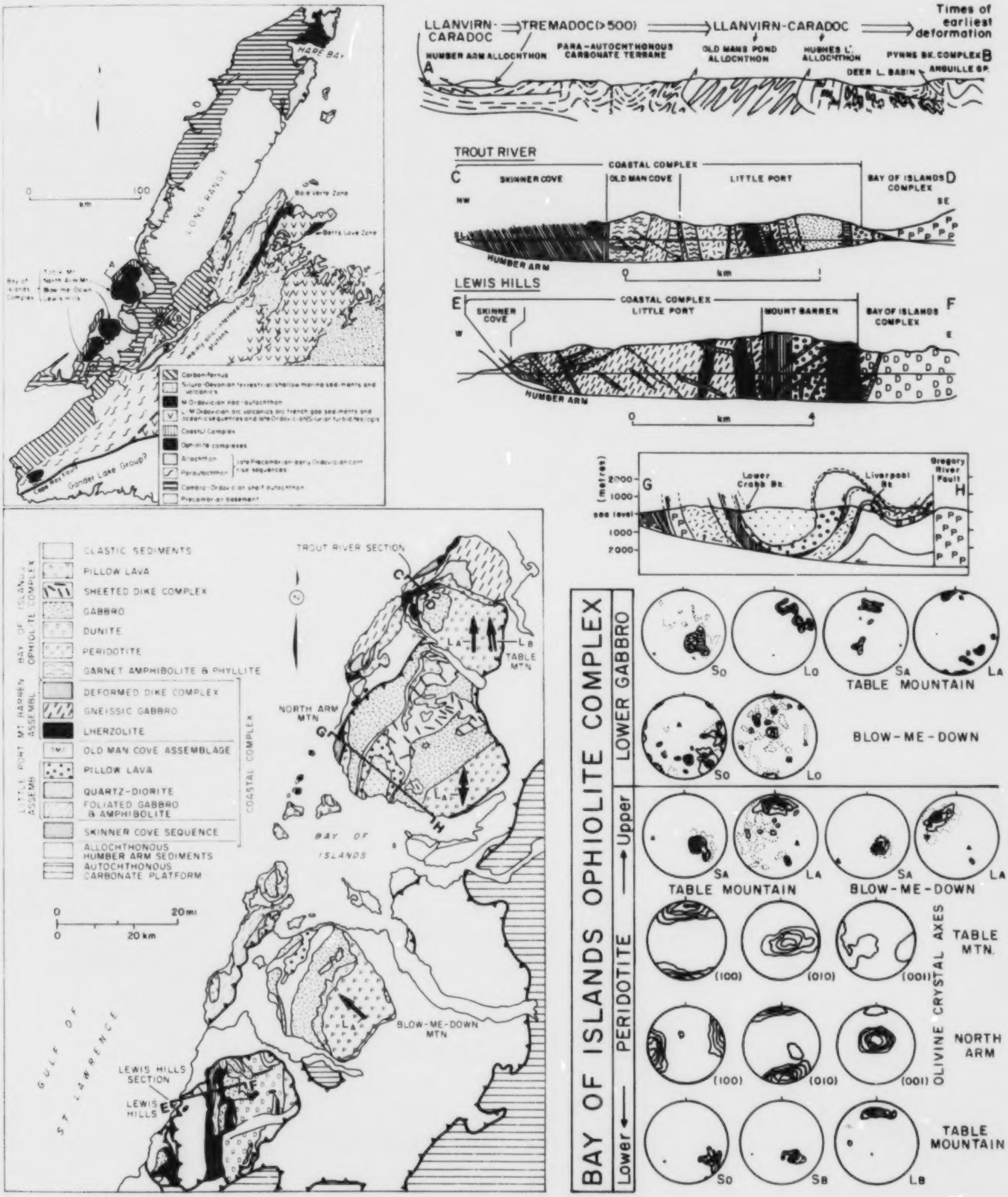

HARE BAY
LONG RANGE
0 100 km
Baie Verte Zone
Betts Cove Zone
A
B
Bay of Islands Complex
Table Mt
North Arm Mt
Blow-me-Down
Lewis Hills
mainly silic - intermediate plutons
Cape Ray Fault
Gander Lake Group?
Carboniferous
Siluro-Devonian terrestrial/shallow marine sediments and volcanics
M Ordovician neo-autochthon
L-M Ordovician arc volcanics arc trench gap sediments and oceanic sequences and late Ordovician/Silurian turbidites/cgls
Coastal Complex
Ophiolite complexes
Allochthon
Parautochthon
late Precambrian-early Ordovician cont rise sequences
Cambro-Ordovician shelf autochthon
Precambrian basement
LLANVIRN-CARADOC
TREMADOC(>500)
LLANVIRN-CARADOC
Times of earliest deformation
HUMBER ARM ALLOCHTHON
PARA-AUTOCHTHONOUS CARBONATE TERRANE
OLD MANS POND ALLOCHTHON
HUGHES L. ALLOCHTHON
PYNNS BK. COMPLEX
DEER L. BASIN
ANGUILLE GP.
TROUT RIVER
COASTAL COMPLEX
C
SKINNER COVE
OLD MAN COVE
LITTLE PORT
BAY OF ISLANDS COMPLEX
D
NW
SE
HUMBER ARM
0 km 1
LEWIS HILLS
E
COASTAL COMPLEX
SKINNER COVE
LITTLE PORT
MOUNT BARREN
BAY OF ISLANDS COMPLEX
F
W
E
HUMBER ARM
0 km 4
(metres)
2000
1000
sea level
1000
2000
G
Lower Crabb Bk.
Liverpool Bk.
Gregory River Fault
H
BAY OF ISLANDS OPHIOLITE COMPLEX
CLASTIC SEDIMENTS
PILLOW LAVA
SHEETED DIKE COMPLEX
GABBRO
DUNITE
PERIDOTITE
GARNET AMPHIBOLITE & PHYLLITE
DEFORMED DIKE COMPLEX
GNEISSIC GABBRO
LHERZOLITE
OLD MAN COVE ASSEMBLAGE
PILLOW LAVA
QUARTZ-DIORITE
FOLIATED GABBRO & AMPHIBOLITE
SKINNER COVE SEQUENCE
ALLOCHTHONOUS HUMBER ARM SEDIMENTS
AUTOCHTHONOUS CARBONATE PLATFORM
MT BARREN ASSEMBL
LITTLE PORT ASSEMBL
COASTAL COMPLEX
0 20 mi
0 20 km
TROUT RIVER SECTION
TABLE MTN
NORTH ARM MTN
BAY OF ISLANDS
BLOW-ME-DOWN MTN
GULF OF ST LAWRENCE
LEWIS HILLS SECTION
LEWIS HILLS
LA
LB
N
BAY OF ISLANDS OPHIOLITE COMPLEX
LOWER GABBRO
Lower ← PERIDOTITE → Upper
So
Lo
SA
LA
TABLE MOUNTAIN
BLOW-ME-DOWN
TABLE MOUNTAIN
BLOW-ME-DOWN
(100)
(010)
(001)
TABLE MTN
NORTH ARM
OLIVINE CRYSTAL AXES
SB
LB
TABLE MOUNTAIN

Figure 14. Map and section of the Humber Arm Old Man's Pond and Hughes Lake Allochthons to show the major structure (after Williams, 1975; Karson & Dewey, 1978; Casey & Kidd, 1981; Williams *et al.*, 1982). Also shown are the fabric elements and dyke complex in the Bay of Islands ophiolite complex (after Girardeau & Niccolad, 1981; Mercier, 1976; Christensen & Salisbury, 1979; Rosencrantz, 1983).

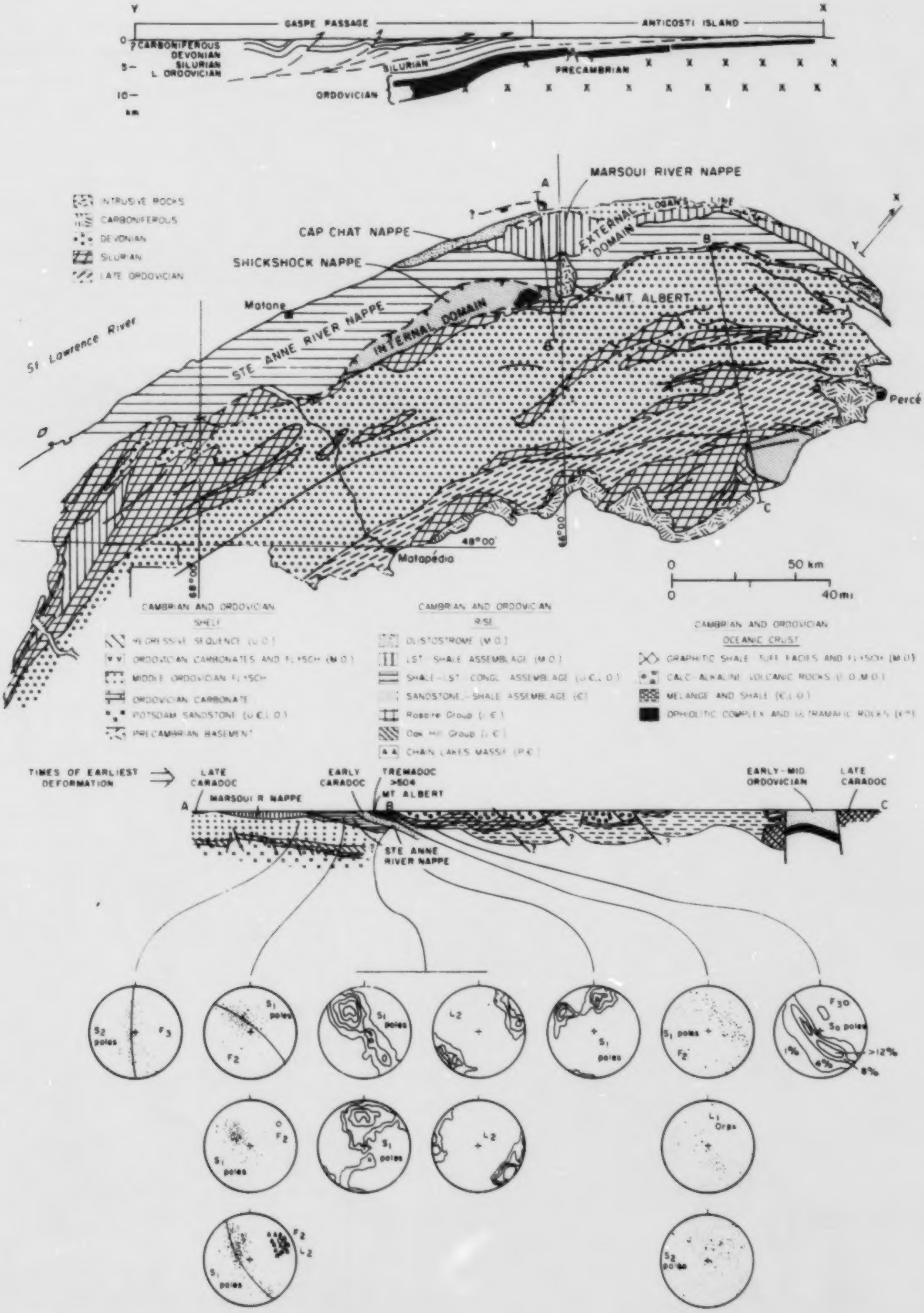

Y
GASPE PASSAGE
ANTICOSTI ISLAND
X
CARBONIFEROUS
DEVONIAN
SILURIAN
L. ORDOVICIAN
SILURIAN
PRECAMBRIAN
ORDOVICIAN
0
5
10
km
INTRUSIVE ROCKS
CARBONIFEROUS
DEVONIAN
SILURIAN
LATE ORDOVICIAN
MARSOUI RIVER NAPPE
CAP CHAT NAPPE
SHICKSHOCK NAPPE
EXTERNAL DOMAIN
LOGAN'S LINE
INTERNAL DOMAIN
STE ANNE RIVER NAPPE
MT. ALBERT
Matane
St Lawrence River
Percé
Matapédia
48°00'
A
B
C
0
50 km
40 mi
CAMBRIAN AND ORDOVICIAN SHELF
REGRESSIVE SEQUENCE (U.O.)
ORDOVICIAN CARBONATES AND FLYSCH (M.O.)
MIDDLE ORDOVICIAN FLYSCH
ORDOVICIAN CARBONATE
POTSDAM SANDSTONE (U.Є.L.O.)
PRECAMBRIAN BASEMENT
CAMBRIAN AND ORDOVICIAN RISE
OLISTOSTROME (M.O.)
LST.-SHALE ASSEMBLAGE (M.O.)
SHALE-LST.-CONGL. ASSEMBLAGE (U.Є., L.O.)
SANDSTONE-SHALE ASSEMBLAGE (Є)
Rosaire Group (L.Є)
Oak Hill Group (L.Є)
CHAIN LAKES MASSIF (P.Є)
CAMBRIAN AND ORDOVICIAN OCEANIC CRUST
GRAPHITIC SHALE, TUFF FACIES AND FLYSCH (M.O.)
CALC-ALKALINE VOLCANIC ROCKS (L.O.,M.O.)
MELANGE AND SHALE (Є,L.O.)
OPHIOLITIC COMPLEX AND ULTRAMAFIC ROCKS (Є?)
TIMES OF EARLIEST DEFORMATION
LATE CARADOC
EARLY CARADOC
TREMADOC >504
MT ALBERT
MARSOUI R NAPPE
STE ANNE RIVER NAPPE
EARLY-MID ORDOVICIAN
LATE CARADOC
S2 poles
F3
S1 poles
F2
L2
S0 poles
F3
1%
4%
8%
>12%
L1 Orgs

Figure 15. Map and section of the structural elements of the Gaspe Peninsula (after St. Julien & Hubert, 1975). Al shown are the fabric elements (after Sikander & Fyson, 1969; Carrara & Fyson, 1973; De Romer, 1976; Beaudi 1980).

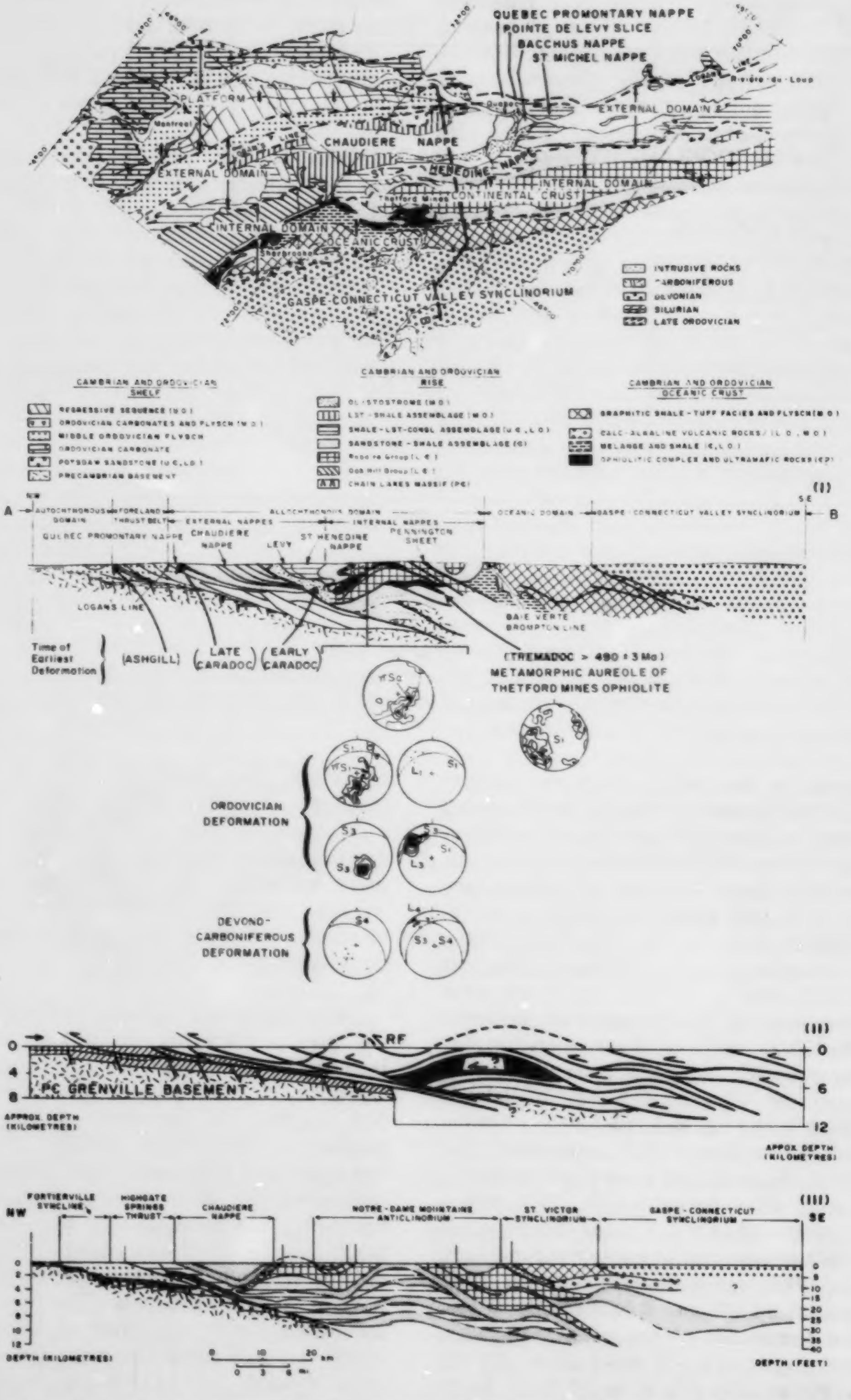

QUEBEC PROMONTARY NAPPE
POINTE DE LEVY SLICE
BACCHUS NAPPE
ST MICHEL NAPPE
Rivière-du-Loup
Quebec
Montreal
PLATFORM
EXTERNAL DOMAIN
CHAUDIERE NAPPE
ST HENEDINE NAPPE
INTERNAL DOMAIN
CONTINENTAL CRUST
Thetford Mines
INTERNAL DOMAIN
OCEANIC CRUST
GASPE-CONNECTICUT VALLEY SYNCLINORIUM
INTRUSIVE ROCKS
CARBONIFEROUS
DEVONIAN
SILURIAN
LATE ORDOVICIAN
CAMBRIAN AND ORDOVICIAN SHELF
REGRESSIVE SEQUENCE (M.O.)
ORDOVICIAN CARBONATES AND FLYSCH (M.O.)
MIDDLE ORDOVICIAN FLYSCH
ORDOVICIAN CARBONATE
POTSDAM SANDSTONE (U.€, L.O.)
PRECAMBRIAN BASEMENT
CAMBRIAN AND ORDOVICIAN RISE
OLISTOSTROME (M.O.)
LST-SHALE ASSEMBLAGE (M.O.)
SHALE-LST-CONGL ASSEMBLAGE (U.€, L.O.)
SANDSTONE-SHALE ASSEMBLAGE (€)
Rosaire Group (L.€)
Oak Hill Group (L.€)
CHAIN LAKES MASSIF (PC)
CAMBRIAN AND ORDOVICIAN OCEANIC CRUST
GRAPHITIC SHALE-TUFF FACIES AND FLYSCH (M.O.)
CALC-ALKALINE VOLCANIC ROCKS (L.O., M.O.)
MELANGE AND SHALE (€, L.O.)
OPHIOLITIC COMPLEX AND ULTRAMAFIC ROCKS (€?)
(I)
A
NW
SE
B
AUTOCHTHONOUS DOMAIN
FORELAND THRUST BELT
ALLOCHTHONOUS DOMAIN
EXTERNAL NAPPES
INTERNAL NAPPES
OCEANIC DOMAIN
GASPE-CONNECTICUT VALLEY SYNCLINORIUM
QUEBEC PROMONTARY NAPPE
CHAUDIERE NAPPE
LEVY
ST HENEDINE NAPPE
PENNINGTON SHEET
LOGANS LINE
BAIE VERTE BROMPTON LINE
Time of Earliest Deformation
(ASHGILL)
(LATE CARADOC)
(EARLY CARADOC)
(TREMADOC > 490 ± 3 Ma)
METAMORPHIC AUREOLE OF THETFORD MINES OPHIOLITE
πS0
S1
πS1
L1
ORDOVICIAN DEFORMATION
S3
L3
DEVONO-CARBONIFEROUS DEFORMATION
S4
L4
(II)
RF
PC GRENVILLE BASEMENT
0
4
8
6
12
APPROX. DEPTH (KILOMETRES)
(III)
NW
SE
FORTIERVILLE SYNCLINE
HIGHGATE SPRINGS THRUST
CHAUDIERE NAPPE
NOTRE-DAME MOUNTAINS ANTICLINORIUM
ST. VICTOR SYNCLINORIUM
GASPE-CONNECTICUT SYNCLINORIUM
0
2
4
6
8
10
12
DEPTH (KILOMETRES)
0
5
10
15
20
25
30
35
40
DEPTH (FEET)
0
10
20
km
0
3
6
mi

Figure 16. Map and sections of the structural elements in southwestern Quebec (after St. Julien and Hubert, 1975, St. Julien *et al.*, 1983; Ando *et al.*, 1983; Laroche, 1983).

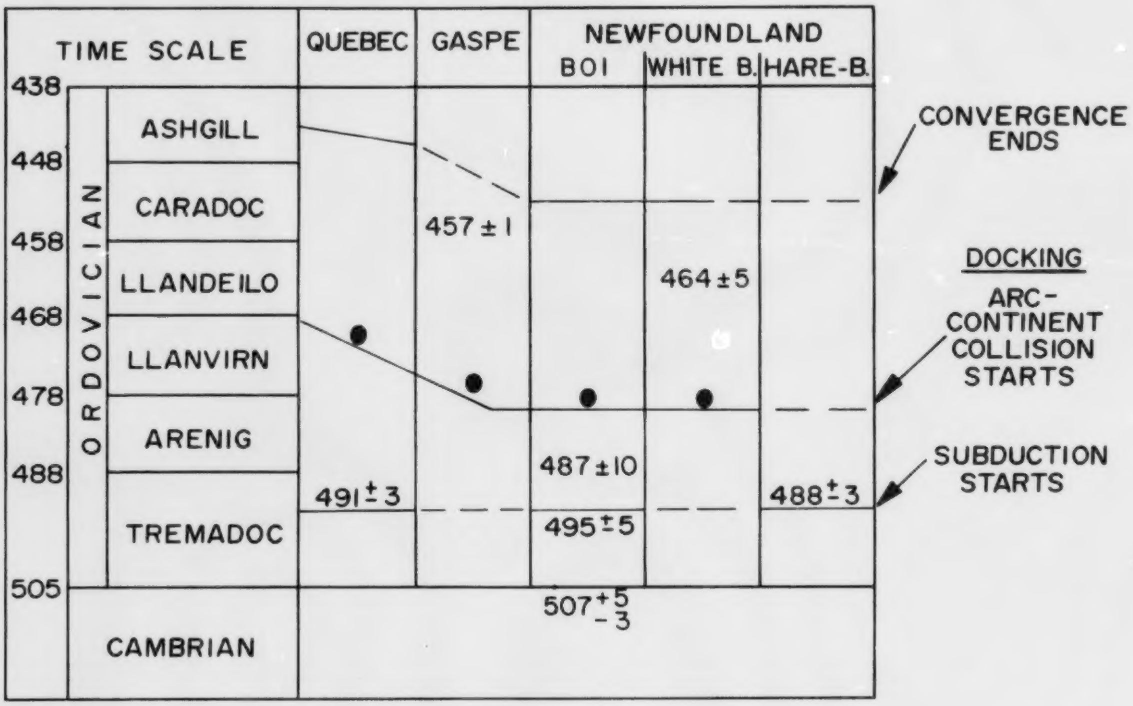

TIME SCALE
QUEBEC
GASPE
NEWFOUNDLAND
BOI
WHITE B.
HARE-B.
438
448
458
468
478
488
505
ORDOVICIAN
ASHGILL
CARADOC
LLANDEILO
LLANVIRN
ARENIG
TREMADOC
CAMBRIAN
457 ± 1
464 ± 5
487 ± 10
491 ± 3
488 ± 3
495 ± 5
507 +5 -3
CONVERGENCE ENDS
DOCKING
ARC-CONTINENT COLLISION STARTS
SUBDUCTION STARTS

Figure 17. Time-Space diagram along the Humber-Notre Dame zones illustrating the synchronous start of subduction and the diachronous start of arc-continental rise collision and the end of convergence.

orthopyroxene in the basal mylonites suggests twophases of deformation, with the larger grains (80 μm) recrystallizing through subgrain rotation (Nicolas and Poirier, 1976) while the smaller grains (10 μm) probably formed through grain boundary migration (Ave Lallemant and Carter, 1970). Jamieson (1986) deduced that the White Hills, basal, high temperature mylonitic peridotites that formed at 850-1050°C and 8.5-10.5 kbar, were locally overprinted by low temperature mylonite under 500-850°C and <3 kbar metamorphic conditions, while McCaig (1983) calculated P-T conditions at the base of the Bay of Islands ophiolite to be 7-11 kbar and 750-850°C. On the other hand, Girardeau (1982) estimated that recrystallation of the 10 μm grain sized mylonitic peridotite took place between 500-750°C and a deviatoric stress of 1.5-2.8 kbar based upon experimental recrystallization under wet conditions (Ross *et al.*, 1980), while the 80 μm grains could have formed at 600°C and 0.05 kbar deviatoric stress. The orientation of the mineral lineation varies about a general NNW trend (Figs. 12-15): (i) NW in White Hills, (ii) N to NNW in Table Mountain, (iii) NNW to NNE in North Arm Mountain, (iv) NNW in Blow-Me-Down Mountain and (v) NNW at Mount Albert. The sense of shear, determined by measuring the angle between the mean extinction of enstatite and the foliation (Nicolas and Poirier, 1976) indicates a NNW movement of the overlying rocks relative to the underlying rocks.

Where a second foliation and mineral lineation are present they are defined by alignment of lamellar enstatite and spinel. In the Bay of Islands Complex, the orientation of this second lineation is 355°/25°N. In the base of the White Hills peridotite, Girardeau (1982) has recorded tight-isoclinal folds and large asymmetrical folds deforming the compositional layering and early foliation. The isoclinal fold axes have a variable trend in the northwest quadrant with a mean northwest trend, whereas the asymmetric fold axes have more variable trends. The isoclinal folds are inferred to have originated as asymmetric folds which suffered tightening and rotation of the fold axes towards the extension direction during progressive deformation. At Glover Island, there is a N-S pressure shadow, stretching lineation (Knapp, 1980). Possibly the same fold generation at Mount Albert has a mean fold axis orientation of 240/45°SW (Beaudin, 1980).

The western part of the Bay of Islands Complex is made up of a steeply dipping, N-S to NNE-SSW, structurally complex zone of sedimentary and igneous rocks (Fig. 14) called the Coastal Complex (Karson and Dewey, 1978). Deformation in the igneous rocks is most complex in the shear zones such as those represented by the Mount Barren and Old Man Cove assemblages. In the Old Man Cove amphibolites, the deformation comprises a mylonitic banding-foliation with a subhorizontal fibre lineation deformed by three generations of folds: tight-isoclinal, upright, steeply plunging folds with an associated foliation, all refolded by gently plunging open folds and later, variably oriented conjugate folds and kink bands. In the Mount Barren assemblage, the deformation consists of (i) an early steeply dipping foliation associated with a gently NW plunging lineation and tight-isoclinal folds deforming the cumulate layering; (ii) boudinage in the foliation and (iii) late high strain zones. These zones pass into the Little Port assemblage of mafic and ultramafic rocks in which deformation varies from a strong to weak foliation and associated gently plunging lineation cut by later high-strain zones and cataclastic zones. Gabbro, diorite, trondhjemite and diabase dykes appear to have been intruded intermittently before, during and after the deformation.

Units Above the Ophiolites

In a fault-bounded segment in the southern part of the North Arm Mountain (and possibly on Blow-Me-Down) massif, all the units of the ophiolite and some thrusts are unconformably overlain by the Crabb Brook Group (Fig. 14) (Casey and Kidd, 1981). This Group consists of breccias (Crabb Point Formation) at the base containing clasts derived from all levels of the ophiolite, metagreywacke and foliated amphibolite clasts derived from the metamorphic sole beneath the ophiolite or from high strain zones within the ophiolite, and rare sedimentary clasts. This unit is overlain conformably by shales and pebbly mudstones of the Jaws Brook Formation which contains Llanvirnian acritarchs. This formation is in turn overlain by sandstones, siltstones and shales of the Summerhouse Brook Formation. Possible correlative breccias in southern Quebec have been assigned to the St. Daniel Formation which rests unconformably upon different units of the ophiolite or conformably above interbedded cherts and volcanic rocks at the top of the ophiolite (Hebert, 1981 and St. Julien, 1980). In Newfoundland, the Crabb Brook Group has been cut by thrusts followed by large scale tight, upright-asymmetric, N-S/NE-SW folds. Elsewhere in the Bay of Islands Complex, similar but larger and gentler folds occur (Williams, 1973). Gravity data over the Bay of Islands Complex indicates that the ophiolite massifs only extend to about 5 km depth (Weaver, 1967). This suggests that the thrust at the base of the complex is subhorizontal and truncates the major late folds (Williams, 1973; Casey and Kidd, 1981). The NW-SE sinistral, strike-slip fault between the North Arm and Table Mountain massifs may be related to this final phase of thrusting.

Units Beneath the Ophiolites

Metamorphic Sole Structures in the metamorphic aureole immediately beneath the ultramafic rocks have generally not received detailed study. In the White Hills area, Girardeau (1982) has recorded that the structures in the basal peridotite and the uppermost metamorphic aureole are parallel. The lineation in the upper aureole consists of aligned amphibole, pyroxene and mineral aggregates, and trends NW-SE. Peak metamorphic conditions of c.900-950°C and 7-10 Kbar are preserved close to the peridotite contact (Jamieson, 1981), while 25 metres beneath the peridotite metamorphic assemblages formed at 860°C and 7-10 kb pressure Jamieson, 1980). These rocks suffered retrograde, epidote amphibolite facies metamorphism in shear zones for which there are no published detailed structural analysis (Jamieson, 1986).

Structures in the rest of the metamorphic sole beneath the ultramafic rocks have not received detailed systematic study. An example of the kind of structures to be expected may be found in Craw (1983), who examined a small part of the greenschist facies sole beneath the White Hills peridotite (Fig. 12). The structures are complex in the Goose Cove Schist with two or more foliations whereas the underlying Ireland Cove Volcanics generally possess only one pervasive foliation. Although Craw (1983) assigned the structures to a sequence of deformational phases all developed under similar metamorphic conditions, it is possible to interpret all these structures in terms of shear zone fabrics, such as C-S fabrics (Berthe *et al*.,

1979) or shape and shear band fabrics (Lister and Snoke, 1984), which may be inhomogeneously developed on both local and regional scales. While the structures in any one shear zone may be approximately synchronous, correlation between shear zones at different structural levels, as inferred by Craw (1983), is improbable because the grade of peak metamorphism decreases gradually downwards from 950-900°C and 7-10 kb pressure immediately beneath the peridotite to 360°C and 2 kb at 700-800 m below the contact (Jamieson, 1980; 1986). This implies erosion or structural thinning of about 20 km during emplacement, which could have taken considerable time. Higher pressures (3-5 kbar) at similar temperatures (300-400°C) have been recorded in the Maiden Point Formation south of Hare Bay (Jamieson, 1986) indicating a considerable difference in the thickness of the tectonic slices along strike. A model of progressive underplating beneath a moving allochthon would also caution one against such correlations. Fold axes and stretching lineations are parallel and E-W trending. This indicates an E-W direction of relative movement in which the folds axes rotated towards the slip direction as shear strain increased probably to values greater than 6 (Escher *et al.*, 1975). It is interesting to note that the E to W movement direction in the greenschist part of the metamorphic sole contrasts with the SE to NW orientation recorded at the top of the sole (Girardeau, 1982). Unfortunately, there are no structural data from the intervening rocks to show whether the change in orientation was gradual or abrupt. Similar structural data are not available from ophiolitic aureoles elsewhere in the Canadian Appalachians.

Sutton-Fleur de Lys Domain (Inner Humber Zone) Along the inner part of the Humber Zone and structurally beneath the ophiolitic and metamorphic sole rocks in Quebec lie metamorphosed pelitic and psammitic rocks of the Sutton-Notre Dame Anticlinorium, Shickshock Mountain and Maquereau Dome (Figs. 15 and 16). Equivalent rocks in Newfoundland occur in the Hughes Lake Allochthon and the Fleur de Lys Supergroup (Figs. 13 and 14). This part of the Humber Zone roughly corresponds to the internal domain of St. Julien and Hubert (1975) characterized by polydeformed rocks metamorphosed at greenschist - amphibolite facies (Figs. 4-8). Although detailed structural studies have been undertaken at various locations in these rocks, they have generally centered on the sequence of structures and their geometry (Figs. 4-8). The structural sequence has usually been correlated throughout the local area and beyond, and the notion that they could represent shear zone structures, which are probably diachronous, was not considered. Also kinematic indicators are rarely recorded. The orientation of early recumbent fold axes are highly variable not only as a result of refolding but presumably due to rotation towards the movement direction. However sheath folds have usually not been recognized. A major WNW-trending sheath fold appears to be present in the Notre Dame Anticlinorium although Charbonneau and St. Julien (1981) interpreted it as a refolded fold. In some places these early folds are associated with thrusts. Later folds tend to be more steeply dipping and trend roughly parallel to the sinuous trend of the orogen verging both towards and away from the North American craton.

Outer Humber Zone Structures in this part of the Humber Zone consist of a series of thrust slices and nappes in continental rise, miogeocline, flysch and basement rocks formed under greenschist to subgreenschist facies metamorphic conditions (Figs. 9, 12-16). Structural complexity grades from polyphase in the southeast through single phase dying out northwestwards. The boundary between inner and outer parts of the Humber Zone is gradational and so is purely arbitrary. In general, the folds in the outer part of the Humber Zone trend parallel or slightly anticlockwise to the strike of the orogen with W-NW-N vergence (locally SE vergence), presumably partly controlled by the Cambrian shape of the North American cratonic margin. It is uncertain how much rotation of the fold axes toward the movement direction has occurred and no kinematic studies of the thrust fault zones have been undertaken. If one assumes that relatively little rotation has taken place then a movement direction somewhere in the northwest quadrant is indicated.

WESTERN DUNNAGE ZONE

In the western part of the Dunnage Zone, Ordovician structures have been recorded in calc-alkaline volcanic rocks of the Ascot-Weedon Formations in the Stoke Mountain area of Quebec and in the ophiolitic rocks east of the Baie Verte Line in Newfoundland (Figs. 4-8).

Stoke Mountain area

Southeast of the St. Victor Synclinorium in

Quebec lies the Stoke Mountain Belt, made up of volcanic rocks of the Ascot-Weedon Formations (Fig. 16). These rocks exhibit polyphase deformation at least some of which is pre-Caradocian because the unconformably overlying, early Caradocian Magog Formation contains pre-depositionally deformed clasts (Lamarche, 1967, 1972; De Romer, 1980, 1981). The earliest structures consist of N-S, W-vergent, recumbent isoclinal folds and thrusts, although the intensity of deformation suggests that the fold axes have probably rotated towards the movement direction. The Moulton Hill granite is presumed to be subvolcanic to the Ascot-Weedon Formations and has yielded a Rb-Sr whole rock errorchron of 505 ± 88 Ma (Poole, 1980).

Newfoundland

East of the Bay Verte-Brompton Line ophiolitic rocks are generally conformably overlain by sediments (including conglomerates containing ophiolitic and pre-depositionally deformed clasts) and calc-alkaline volcanic rocks such as the St. Daniel, Big Head/Barry - Cunningham, Glover Island Formations and Snooks Arm Group (Figs. 4-8). The latter two units contain Areniginan fossils. Structures in the ophiolites are usually confined to a serpentinized zone at the base and at Glover Island are associated with a N-S stretching lineation. Other structures in these rocks appear to affect younger rocks and are generally assigned to Late Paleozoic deformational events.

AGE OF THE STRUCTURES

Palaeontogical and isotopic data provide some constraints upon the ages of these structures. In view of the potential for diachronism, accurate location of sample sites is essential. Unfortunately, such information is not usually reported in the literature, thereby precluding a detailed analysis. Thus, only general trends may be derived here. These data will be considered systematically starting with the ophiolites. A lower limit on the age of the structures in the ophiolite is provided by its crystallization age, namely (i) two Sm-Nd mineral - whole rock isochrons of 508 ± 6 Ma and 501 ± 13 Ma from the pyroxene gabbro of the Bay of Islands (Jacobsen and Wasserburg, 1979); (ii) U-Pb zircon analyses from trondjhemites yielded 504 ± 10 Ma (upper intercept) and 440 ± 30 Ma (lower intercept)(Mattinson, 1976) and 486 ± 2 Ma (abrasion technique, Dunning and Krogh, 1985) in the Bay of Islands Complex; 508 ± 5 Ma (Mattinson, 1975) in the Coastal Complex; 474 ± 2 Ma for tonalite in the Coney Head Complex (U-Pb on zircon; Dunning, 1987a); and 489 ± 3 (-2) Ma (abrasion technique, Dunning and Krogh, 1985) from gabbro in the Betts Cove Ophiolite Complex.

^{40}Ar/^{39}Ar incremental plateau cooling ages from hornblende in (i) undeformed trondjhemite and pegmatitic hornblende gabbro in the Coastal Complex are 495 ± 5 Ma, 492 ± 11 Ma and 499 ± 10 Ma respectively and (ii) highly deformed amphibolites in the Coastal Complex are 459 ± 5 Ma and 464 ± 7 Ma (Idleman, 1985). Although Karson and Dewey (1978) concluded that the structures in the Coastal Complex predated the undeformed trondjhemite bodies, i.e. they were pre-503 Ma, Idleman (1985) believed that time of deformation is bracketed by the ages from undeformed and deformed parts of the Coastal Complex, i.e. between 501-457 Ma. A sample from finely banded mylonitic rocks within the basal ultramafic rocks of the North Arm Mountain, Bay of Islands Complex, has yielded an ^{40}Ar/^{39}Ar incremental release plateau cooling age on hornblende of 487 ± 10 Ma (recalculated from Dallmeyer and Williams, 1975) and places a lower limit on the mylonization which took place at temperatures greater than ca. 500°C. Hornblende from the sheeted dykes in North Arm Mountain have yielded an ^{40}Ar/^{39}Ar incremental release plateau cooling ages of 468 ± 12 Ma (recalculated from Archibald and Farrar, 1976). Other ages from the Bay of Islands/Coastal Complexes are of dubious value because they are K-Ar ages and ^{40}Ar/^{39}Ar incremental release patterns show several components. Nevertheless, they appear to be generally consistent with the ^{40}Ar/^{39}Ar data. Thus K-Ar ages from hornblendes of the gabbro and sheeted dykes of North Arm and Table Mountain massifs range from 451 to 470 ± 13 Ma and from foliated amphibolites in the Coastal Complex range from 466 to 486 ± 13 Ma (recalculated from Archibald and Farrar, 1976).

Isotopic data from other ophiolitic complexes are rather limited and of uncertain value. Thus, a granitic body intrusive into the Thetford Mines ophiolite yielded a Rb-Sr whole rock isochron of 466 ± 13 Ma (Poole, 1980). However, even though it is deformed, its position in the structural history of the ophiolite was not reported. Also, Laurent (1977) interpreted these granites as tectonic slivers, in which case their age provides no constraints on the age of the ophiolite.

An upper age limit on structures forming at or above ca. 500°C is given by ^{40}Ar/^{39}Ar plateau cooling ages on hornblende in the upper, amphibolitic, metamorphic aureole beneath the ultramafic rocks which have yielded: (i) 482 to 495 ± 10 Ma at White Hills (recalculated from Dallmeyer, 1977); (ii) 467 to 477 ± 10 Ma at North Arm Mountain (recalculated from Dallmeyer and Williams, 1975); (iii) 457 ± 1 Ma and 454 ± 2 Ma at Mont Albert (Lux, 1985); and (iv) 491 ± 3 Ma (Clague *et al.*, 1981) and 482 Ma (Lux, 1984) at Thetford Mines. K-Ar ages, of dubious value, from hornblendes in the metamorphic aureole beneath (i) the Bay of Islands range from 467 to 558 ± 16 Ma (recalculated from Archibald and Farrar, 1976) and (ii) Mount Albert gave 485 ± 45 Ma (Wanless *et al.*, Lowdon *et al.*, 1962; Wanless *et al.*, 1973) and probably reflect a multicomponent ^{40}Ar/^{39}Ar incremental release patterns.

Sedimentological/paleontological constraints upon uplift and erosion, probably associated with the overthrusting of the ophiolite are provided by: (a) the first appearance of easterly derived ophiolitic detritus in the allochthonous middle Arenigian (I. gibberulus Zone) flysch of the Eagle and Lower Head sandstones (formerly included in the Blow-Me-Down Brook Formation), central western Newfoundland (Stevens, 1970, 1976) and in late Arenigian Tourelle, Metis and St.-Modeste Formations, Gaspe (Hiscott, 1978); (b) foliated and unfoliated clasts of diabase and basalt in the (?)Early Ordovician Skinner Cove Formation (Williams, 1975); (c) the angular unconformity beneath the Llanvirnian Crabb Brook Group which cuts across all units of the North Arm Mountain ophiolite (Casey and Kidd, 1981); (d) the occurrence of ophiolitic detritus, some of which are foliated in units conformably overlying the ophiolites, such as, in the middle-late Arenigian Glover Island Formation, in the undated Baie Verte Group, in the Big Head and St. Daniel Formations both of unknown age (Knapp, 1980; Riley, 1957; Kidd, 1977; Kidd *et al.*, 1978; St. Julien, 1980; Williams and St. Julien, 1982); and (e) the presence of deformed metasedimentary clasts in units conformably overlying the ophiolites, such as, the Arenigian Snooks Arm Group, the undated Baie Verte Group and the cherty argillite lying above the Thetford Mines Ophiolite (Upadhyay *et al.*, 1971; Upadhyay, 1973; DeGrace *et al.*, 1976; Kidd, 1977 and Laurent, 1975, 1977). Note that the units above the ophiolite are conformable east of the Bay Verte-Brampton Line and are only unconformable farther west in the Humber Zone, suggesting that uplift was local rather than regional. Also note that the conformable relationship cannot be used to determine whether or not deformation was occurring at these localities because deformation could have been taking place below sea level and at depth, or alternatively, may have been absent with the deformed clasts being imported.

Considering all the foregoing isotopic and stratigraphic constraints and using Palmer's (1983) time scale, an upper limit for the onset of deformation associated with convergence is (i) pre-491 ± 3 Ma at Thetford Mines; (ii) pre-Late Arenig (i.e. pre-480 ± 20 Ma) in the Gaspe; (iii) pre-487 ± 10 Ma or pre-middle Arenig (i.e. pre-485 ± 20 Ma) in the Bay of Islands; and (iv) pre-495 ± 10 Ma in the White Hills. A lower limit of 508 ± 6 Ma is only available in the Bay of Islands. Thus, within the error limits, the onset of Taconian obduction appears to have started synchronously during the Tremadocian in an oceanic environment all along the western side of the Canadian Appalachians.

The timing of subsequent convergence of ancient North America and the deformation front is mainly recorded in the stratigraphic record because no attempt has yet been made to systematically date cleavages between the ophiolite base and the autochthon. These stratigraphic constraints generally appear as black graptolitic shale overlain by easterly-derived wacke and sandstone representing distal and proximal flysch, respectively. The first appearance of these units in the different facies belts bordering ancient North America records the convergence in a general way. Deposition of the flysch was accompanied by normal faulting, e.g. Nicolet Formation (Walters *et al.*, 1982). East of the Taconian front, deposition of this flysch is followed by thrust transport, which ultimately terminates with the end of convergence. In the Canadian Appalachians, these events were diachronous both across strike from southeast (older) to northwest (younger) and also along strike from northeast (older) to southwest (younger) as the following data will show (Figs. 4-8, 17). Thus, the first appearance of flysch in the carbonate bank edge conglomerate facies formed at the base of the continental slope is latest Arenigian-earliest Llanvirnian in the Cow Head Group of Newfoundland (Stevens, 1970; Kindle and Whittington, 1958; James and Stevens, 1986), late Arenigian in the Tourelle Formation lying unconformably upon Cap des Rosiers Group in the Gaspe (Hiscott, 1978) and Llanvirn in the Riviere Ouelle Formation near Quebec City (Hubert, 1973; St. Julien and Hubert, 1975). This stage is also recorded on the continental shelf by the disconformity of late Arenigian-early Llanvirn age between the St. George and Table Head Groups in western Newfoundland (Whittington and Kindle, 1963; Klappa *et al*., 1980). The first appearance of distal flysch on the outer part of the continental shelf is early Llanvirn in the black shales of the Hare Island Limestone (= middle Table Head Group) at Hare Bay, northern Newfoundland (Fahraeus, 1970; Stevens, 1970), whereas it is earliest Caradocian in both the Deslandes Formation of the Gaspe and the Bullstrode Formation near Quebec City (St. Julien and Hubert, 1975; Riva, 1974). Along the inner part of the continental shelf, this same distal flysch is middle-late Llanvirnian (Diplograptus decoratus Zone) along the western coast of Newfoundland (Klappa *et al*., 1980; Finney and Skevington, 1979) and middle Caradocian in the Cloridorme Formation of Gaspe (Enos, 1969) and in the Utica Shale, Beaupre and St.-Irenee Formations near Quebec City (Riva, 1974; Belt *et al*., 1979; Belt and Bussieres, 1981). The proximal flysch (i) is Llanvirnian in the Goose Tickle Formation at Hare Bay, northern Newfoundland (Stevens, 1970; Erdtmann, 1971); (ii) in the Mainland Sandstone of central western Newfoundland, it varies in age from Llanvirnian in the east to early Caradocian (D. multidens Zone) in the west (Stevens, 1970, 1976; Bergstrom, *et al*., 1974; Rogers and Neale, 1963); (iii) is middle-late Caradocian in the Cloridorme Formation of Gaspe (Enos, 1969) and (iv) near Quebec City and Montreal, it varies from earliest Caradocian (N. gracilis Zone) in the Etchemin River Olistostrome in the southeast to early Caradocian-earliest Ashgillian in the Nicolet River Formation in the North (St. Julien and Hubert, 1975; Riva, 1972, 1974; Beaulieu *et al*., 1980; Walters *et al*., 1982).

The age constraints on the final stages of Taconian thrust deformation vary from middle Caradocian in Newfoundland to middle Ashgillian in Quebec. In Newfoundland, the late Llanvirnian Goose Tickle Formation was eventually overridden by the Hare Bay Allochthon but no upper age constraint is available (Stevens, 1970; Erdtmann, 1971). In the Bay of Islands the Llanvirnian Crabb Brook was involved in westward thrusting and folding. An upper limit of 459 ± 5 Ma to 467 ± 10 Ma (Llanvirn to early Caradocian) for deformation taking place above ca. 500°C is provided by the youngest ^{40}Ar/^{39}Ar incremental release plateau ages on hornblende from the Coastal Complex and North Arm Mountain, respectively (Idleman, 1985; Dallmeyer and Williams, 1975). An upper limit on lower temperature deformation is only available farther south on the Port au Port Peninsula, where the age of final emplacement of the toe of the Humber Arm Allochthon is precisely dated as middle-Caradocian (D. multidens Zone), the age not only of the youngest part of the Mainland Sandstone west of the allochthon but also of the Long Point Formation which unconformably overlies the allochthon (Rogers and Neale, 1963; Bergstrom *et al*., 1974 and Stevens, 1976). However, there could be other younger thrusts at deeper levels which may be blind or surface in the Gulf of St. Lawrence west of Newfoundland (Cawood and Williams, 1988). This possibility could find some support if any of the folding along the west coast of Newfoundland is Ordovician in age. These folds are presently assigned to the Late Paleozoic because Siluro-Devonian rocks are also affected (Keppie *et al*., 1982).

Farther east in the Humber Zone of Newfoundland, it becomes more difficult to separate the effects of Ordovician from Late Paleozoic deformation. The distinction may be made in the Southern White Bay Allochthon where the Middle-Silurian Sops Arm Group rests unconformably upon the Coney Head Complex in some places, but elsewhere the Sops Arm Group has been overridden by a thrust slice containing Coney Head Complex (Fig. 13) (Smyth and Schillereff, 1982). An upper limit for the polyphase deformation taking place above ca. 500°C in the western Fleur de Lys is provided by the oldest ^{40}Ar/^{39}Ar incremental release plateau cooling ages on hornblende of 428 ± 10 Ma (recalculated from Dallmeyer, 1977). The Dunamagon Granite, while exhibiting some deformation, cuts across the Baie Verte Line and thereby provides an upper limit on the thrust contact between the Dunnage and Humber Zones. It has yielded a U-Pb zircon age of 459 +27/-15 Ma (Dallmeyer and Hibbard, 1984).

In the Gaspe, an upper limit for the emplacement of the Taconian allochthons is provided by the age of the oldest rocks lying unconformably above them, namely the Ashgillian-(?)latest Caradoc (D. clingani Zone) - middle Llandoverian Cabano Formation (David *et al*., 1985). The $^{40}Ar/^{39}Ar$ incremental release plateau age of 454 ± 24 Ma (Lux, 1985) on hornblende provide an upper limit on the deformation taking place above ca. 500°C. The K-Ar ages on mica range from 504 to 449 ± 24 Ma (recalculated from Lowdon *et al*., 1962 and Carrara and Fyson, 1973) and while they undoubtedly reflect multiple components, the youngest date is generally consistent with the stratigraphic constraint. Farther southeast, in the Gaspe Synclinorium, the middle Caradocian (C. spiniferis) age of the lower part of the Honorat Group (Riva, 1981; Lesperance *et al*., 1981) and the Llandovery A age of the basal Chaleur Group (Bourque and Lachambre, 1980) provide an upper limit on the Taconian structures. This is the same age as the upper part of the Cloridorme Formation on the northern coast of the Gaspe (Enos, 1969) but here it provides a lower limit on the Taconian structures. Thus, some south to north diachronism in the end of the Taconian deformation is indicated.

In the Montreal-Quebec City area, the final emplacement of the Taconian allochthons is recorded where the synorogenic clastic flysch of the Nicolet River Formation changes to post-tectonic limestones and deltaic redbeds in the Becancour River Formation during the late Ashgillian (Beaulieu *et al*., 1980; Clark, 1964). In the Notre Dame Anticlinorium, K-Ar dates on muscovite of 449 ± 15 Ma (recalculated from Lowdon *et al*., 1962 and Wanless *et al*., 1979) are at least consistent with this stratigraphic data. Some deformation in the Ascot-Weedon Formations predates the unconformably overlying Magog Group of early Caradocian age but the lower limit is presently very poorly constrained at an unreliable 505 ± 88 Ma age (Poole, 1980).

Thus, in conclusion, it is clear that the Taconian deformation began synchronously in oceanic lithosphere along the western side of the Canadian Appalachians during the Tremadocian soon after the formation of the oceanic lithosphere. The onset and termination of Taconian deformation are progressively younger across strike towards the northwest across the margin of ancient North America. The Taconian deformation appears to have terminated diachronously along strike ending during the early Carodocian in Newfoundland and during the middle Ashgillian in Quebec (Fig. 17). However, it is important to note that the along-strike diachronism can only be documented between Newfoundland and Quebec and within each Province alone events appear to be synchronous along strike.

INTERPRETATION

Internal structures in the ophiolites obducted along the western side of the Canadian Appalachians have usually been interpreted in terms of processes occurring during the formation of oceanic lithosphere at a transform faulted mid-oceanic ridge (Dewey and Kidd, 1977; Girardeau and Nicholas, 1981) and only rarely have they been attributed to the obduction process (Girardeau, 1982). In the mid-oceanic ridge hypothesis, layering in the gabbros is related to either plating in the upper gabbros or magmatic accumulation in the lower gabbros. The foliation and mineral lineation in the ultramafic rocks were inferred to be produced by the relative motion of the rising asthenosphere and the separating lithosphere at the mid-oceanic ridge. In this case, one would expect the temperature of foliation development to increase with depth. Geometrically, the cumulate banding in the lower gabbros dips towards the ridge, whereas, the foliation in the ultramafics dips away from the ridge with the mineral lineation forming parallel to the relative plastic flow direction and generally perpendicular to the ridge. Structures in the Coastal Complex were interpreted in terms of the processes operating in a transform-fracture zone offsetting a mid-oceanic ridge (Karson and Dewey, 1978). Kinematic modelling of the transform- fracture zone shows that it passes from a generally strike-slip regime in the transform segment to dip-slip in the non-transform segment. The deformation is asymmetric in which one strip with a transform fault deformation history is welded to younger crust with no transform history. Geometrically, the transform-fracture zone would form perpendicular to the mid-ocean ridge and the sheeted dykes. On the other hand, Dunsworth *et al*.(1988) have interpreted the structures in the Mount Barren part of the Coastal Complex in terms of spreading related mega-shear zones. Significantly, few if any structures in the ophiolites were attributed to obduction processes by the proponents of these hypotheses.

However, several contradictions arise from the application of this transform-faulted, mid-oceanic ridge model to the obducted ophiolites along the western side of the Canadian Appalachians. Thus the sheeted dykes in North Arm Mountain vary about a mean N-S orientation (Rosencrantz, 1983) and are roughly parallel (when they should be perpendicular) to the NNE- to NNW- trending mineral lineation and NNE-trending Coastal Complex "transform zone". On the other hand, on Blow-Me-Down Mountain, the NW-SE trending sheeted dykes (Williams and Malpas, 1972) are roughly perpendicular to the NNE trending Coastal Complex as predicted by the model, however, the mineral lineation still has an anomalous NNW trend. Similarly, the N- to NNW-trending mineral lineation in Table Mountain is oblique rather than perpendicular to the NNE-trending Coastal Complex. These geometric anomalies may be accounted for if the Coastal Complex is a spreading related mega-shear as proposed by Dunsworth *et al.*(1988). Rotation of the structures adjacent to a transform zone or during obduction could also account for these geometric anomalies, however, other data suggest that at least some of the structures in the ultramafics formed during obduction. Thus, estimates of the temperatures under which the foliation in the ultramafic formed, show a decrease with depth (Girardeau and Nicholas, 1981; Girardeau, 1982) rather than increase as predicted by the mid-ocean ridge model. Furthermore, the 950-900°C temperature and 7-10 Kb pressure estimated for the peak metamorphic conditions just beneath the White Hills obducted ophiolite (Jamieson, 1986) indicate that initial obduction was a hot process quite capable of producing a foliation and lineation of pyroxene, plagioclase, spinel and chromite. This is also compatible with the 500-750°C temperature estimates for development of the olivine - orthopyroxene mylonites in the basal ultramafics (Girardeau, 1982) although these are probably not accurate absolute values (Poirier and Guillope, 1979). Such a conclusion would appear to be supported by the parallelism of linear structures in the White Hills basal ultramafics and the uppermost metamorphic sole, although rotation of older structures towards the slip direction cannot be entirely discounted. Thus, it would appear that obduction initially took place in a NNW direction. In view of these ideas, the thought arises that some of the structures in the Coastal Complex could also have formed during obduction as a lateral ramp zone, although, once again, one cannot entirely discount the possibility that it nucleated on a transform fault. Another possible consequence of such hot obduction is the potential for magma generation which could produce the complex relationships between igneous intrusions, such as trondjhemite and mafic bodies, and the deformation (Karson and Dewey, 1978). Thus, the relatively late 486 ± 2 Ma trondjhemite in the Bay of Islands (Dunning and Krogh, 1985) could have been emplaced during or soon after the initiation of obduction.

Age data in the Bay of Islands Complex and its metamorphic sole indicate that the initiation of obduction could have begun as soon as 5 Ma after the formation of the oceanic lithosphere (Fig.17), i.e. while it was still young, hot and relatively elevated. This raises the possibility that subduction nucleated either on the mid-oceanic ridge if it trended parallel to the N-S sheeted dykes on North Arm Mountain or on a transform zone if the ridge was parallel to the NW-SE sheeted dykes on Blow-Me-Down Mountain. Linear structures in the uppermost part of the metamorphic sole and basal ultramafics indicate that the initial Tremadocian slip direction was towards NNW (Fig. 18). The 7-10 Kbar pressure estimate for the White Hills uppermost aureole/basal peridotite (Jamieson, 1980) indicates that the oceanic lithosphere obducted initially was 28 ± 5 km thick. By Arenigian times, obduction had probably uplifted the ophiolite above sea level thereby providing a source for the ophiolitic detritus (Fig. 18). Such uplift and erosion led to cooling below ca. 500°C of the ophiolite and its amphibolite facies metamorphic sole by about 490 Ma. It should be noted that some of this erosion could have been by tectonic processes (structural thinning) as well as subaerial weathering. By the time about 20-25 km of erosion or structural thinning had taken place and metamorphic conditions beneath the obducted ophiolite were about 360°C and 2 Kbar at 700-800 m below the base of the White Hills ophiolite (Jamieson, 1980; 1986), the slip direction had become due westward (Fig. 18). Whether the change in slip direction from NNW to W was abrupt or gradual is presently unknown. This change in the slip direction may have been due to a change in the convergence direction across the subduction zone. Limited kinematic indicators suggest that the final stages of obduction may have been towards the northwest, although the influence of the trend of the Cambrian continental margin is difficult to distinguish.

The NNW-W polarity of the obduction together with the W-NNW diachronism of the onset of deformation indicate that the subduction zone off ancient North America dipped towards the southeast

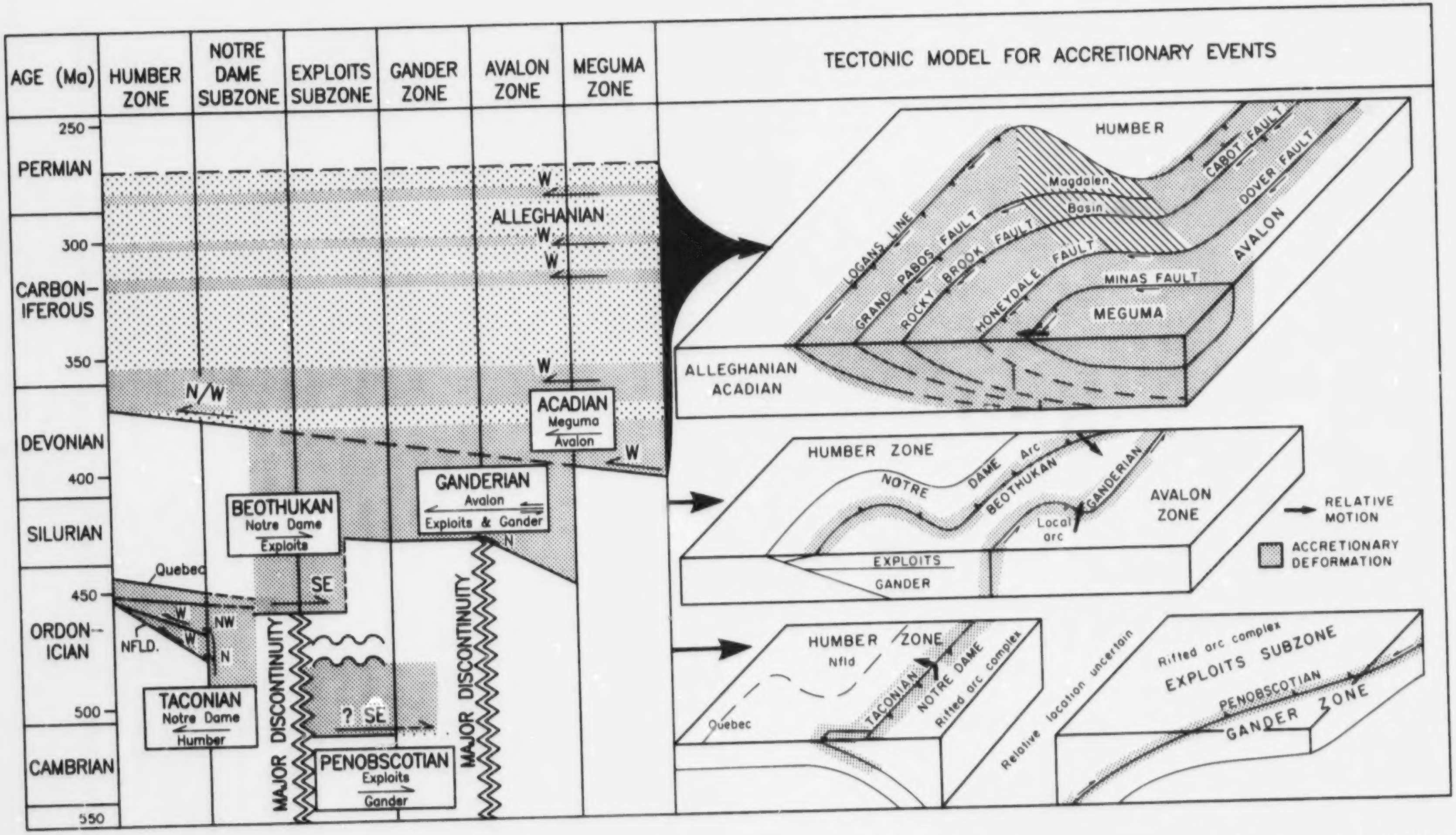

AGE (Ma)
HUMBER ZONE
NOTRE DAME SUBZONE
EXPLOITS SUBZONE
GANDER ZONE
AVALON ZONE
MEGUMA ZONE
TECTONIC MODEL FOR ACCRETIONARY EVENTS
250
PERMIAN
300
CARBON-IFEROUS
350
DEVONIAN
400
SILURIAN
450
ORDON-ICIAN
500
CAMBRIAN
550
ALLEGHANIAN
W
N/W
ACADIAN
Meguma
Avalon
GANDERIAN
Avalon
Exploits & Gander
N
BEOTHUKAN
Notre Dame
Exploits
SE
Quebec
NFLD.
NW
N
TACONIAN
Notre Dame
Humber
MAJOR DISCONTINUITY
? SE
PENOBSCOTIAN
Exploits
Gander
ALLEGHANIAN
ACADIAN
HUMBER
Magdalen
Basin
LOGANS LINE
GRAND PABOS FAULT
ROCKY BROOK FAULT
HONEYDALE FAULT
MINAS FAULT
CABOT FAULT
DOVER FAULT
AVALON
MEGUMA
HUMBER ZONE
NOTRE DAME Arc
BEOTHUKAN
GANDERIAN
Local arc
AVALON ZONE
EXPLOITS
GANDER
RELATIVE MOTION
ACCRETIONARY DEFORMATION
HUMBER ZONE
Nfld
Quebec
TACONIAN
NOTRE DAME
Rifted arc complex
Relative location uncertain
Rifted arc complex
EXPLOITS SUBZONE
PENOBSCOTIAN
GANDER ZONE

Figure 18. Time-Space diagram for accretionary deformation relative to zones in the Canadian Appalachians, accompanied by tectonic models to explain the accretionary events.

(Fig. 18). This results from the observation that obduction and subduction are mutually opposite relative terms analogous to overthrusting and underthrusting. Although it is clear that continued subduction led to progressively more widespread and deeper levels of accretionary deformation and gave rise to the model of progressive underplating, more commonly called the "piggyback" model, it is also clear the deformation was not restricted to one level at a time. Thus, deformation of the Llanvirnian Crabb Brook Group resting above the ophiolite was taking place concurrently with deformation at a deeper level. On the other hand, a deverticulation or unpeeling model of emplacement has also been proposed in which progressive unroofing of an allochthon by gravity sliding produces the opposite sequence with younger slices lying above older ones (St. Julien and Hubert, 1975; Stanley and Ratcliffe, 1983). However, a gravity sliding mechanism would appear to be impossible because the emplacement of an assembled allochthon on the continental margin would cause a lithospheric downwarp (Beaumont, 1981; Quinlan and Beaumont, 1984) producing a foreland basin and a dip on the thrusts towards the ocean, i.e. opposite to that required for gravity sliding. Thus, it is proposed that a piggyback model modified by active thrusting at various levels in the allochthonous rocks best explains the observed relationships. Another result of loading the continental margin with allochthons is the development of a flexural bulge which was recorded in the late Arenigian-early Llanvirian disconformity between the St. George and Table Head Groups in western Newfoundland.

The rate of convergence during the Taconian in the Canadian Appalachians requires knowledge of the distance telescoped in a given time interval. Methods restricted to an undeformed foreland basin cannot be applied here. Unfortunately, at present, balanced cross sections have not been constructed across any part of the Canadian Appalachians, so at best only crude minimum estimates can be made. No estimates can be made about the initial N-S convergence rate because the amount of subducted oceanic lithosphere is unknown. A crude estimate of 350 km for the minimum width for the continental shelf, slope and rise which were thrust beneath the Newfoundland ophiolites may be derived by adding to the presently exposed width of the Humber Zone (100 km), the width of the allochthonous rocks of the continental rise and slope (75 km) and a conservative 50% additional shortening as internal strain, folding and thrusting across the Humber Zone and allochthonous slope/rise (175 km). This total shortening amounts to 70% and is comparable with the 80% shortening derived by Dewey (1982) for the Grampian deformation in the British Isles. The 350 km is comparable with the 450 ± 100 km between the edge of the flexural bulge and the deformation front predicted for theoretical models of the Southern Appalachians (Quinlan and Beaumont, 1984). In westernmost Newfoundland, the disconformity between St. George and Table Head Groups is attributed to such flexural bulging. It is also comparable with the distance of 400 km derived by Williams (1980). Thus, in Newfoundland, the Bay of Islands Complex travelled at least 350 km between early Llanvirnian and middle Caradocian, a period of c.25 Ma. The 450 ± 100 km value gives a rate of 1.8 ± 0.4 cm/year for WNW-ESE convergence or 2 ± 0.4 cm/year for W-E convergence. Similar calculations in Quebec give a minimum width of 200 km overridden in 15 Ma between basal Caradocian and middle Ashgillian for a minimum rate of 1.3 cm/year for NW-SE convergence or 2 cm/year for W-E convergence. These convergence rates compare favorably with the 2-3 cm/year rates derived for the New England Appalachians (Cisne, 1982; Bradley *et al.*, 1985).

An alternative method of calculating the convergence rate involves the along-strike diachronism observed in arrival of the deformation front at the edge of the continental shelf: basal Llanvirnian in Newfoundland and basal Caradocian in Quebec, a difference of c.20 Ma (Fig. 17). In this method, the strike of the subduction zone is assumed to be NE-SW parallel to the Cambrian margin of North America. It is also assumed that no transform faults offset the trench. A small circle subduction zone trending roughly parallel to, and concave towards, the ancient North American continental margin finds some support in the apparent lack of along-strike diachronism in the appearance of Llanvirnian distal flysch on the continental margin between southernmost Appalachians and Newfoundland (Hiscott, 1984) and within Newfoundland or Quebec alone. Using the values derived earlier for the restored edge of the continental shelf, the offset between Newfoundland and Quebec would be c.600 km. This gives a rate of 3.0 cm/year for NW-SE convergence and 4.8 cm/year for W-E convergence. These rates are higher than those derived for the

New England Appalachians suggesting that transform faults probably offset the trench (Fig. 18).

After the arrival of the ancient North American continental margin at the trench, the miogeocline and the Grenvillian basement became involved in the thrust deformation. The buoyancy of this continental lithosphere eventually choked the subduction zone, although it took 25 Ma in Newfoundland and 15 Ma in Quebec (Fig. 17). The nappe loading of the outer edge of the ancient continental margin depressed the lithosphere producing the medium pressure amphibolite facies metamorphism and accompanying deformation seen in the Fleur de Lys Supergroup and correlatives. Progressive loading eventually leads to oversteepening of the subducting lithosphere (Stockmal *et al.*, 1986) and could produce the backfolding phenomenon in which folds verge towards the trench. The collision of ancient North America with the trench probably also led to deformation in the volcanic arc complex as is observed in the pre-Caradocian deformation of the Ascot-Weedon Formations and possibly elsewhere. The earlier termination of Taconian deformation in Newfoundland compared with Quebec suggests that a WNW-trending transform fault existed between these segments allowing subduction to continue independently in Quebec. Such a transform fault would probably be parallel to the relative convergence direction and its orientation is consistent with the slip directions derived from the structures. Convergence may also have continued in Newfoundland if lithospheric delamination is invoked leading to tectonic wedging of the lower crust and blind subduction (Stockmal *et al.*, 1987; Cawood and Williams, 1988). The end of subduction led to isostatic uplift and erosion of the continent-volcanic arc collision zone and its roots were then submerged beneath the Late Ordovician-Silurian sea.

The pre-Caradocian deformation recorded in the calc-alkaline volcanic arc in the western Dunnage Zone may be inferred to result from the continent-arc collision. During such a collision the volcanic arc would be thrust over the continent.

Early Paleozoic Structures in the Dunnage (East of the Axis of Vergence Reversal), Gander and Avalon Zones

The central and eastern parts of Newfoundland have been subdivided into several zones and subzones separated by major faults: NOTRE DAME AND DASHWOOD SUBZONES-Red Indian Fault-EXPLOITS SUBZONE-Noel Pauls/Grub Fault-GANDER/MEELPAEG/ MOUNT CORMACK ZONE-Bay d'Est/Dragon Bay Fault-BURGEO TERRANE-Dover/ Hermitage Fault-AVALON ZONE (Williams *et al.*, 1988, 1989; Keppie and Dallmeyer, 1989; Currie and Paisecki, 1989; Piasecki *et al.*, 1990). Similar units in New Brunswick are: Deveraux Formation in the Elmtree inlier (= Notre Dame), the upper part of the Tetagouche Group in the Mirimichi Highlands and the rest of the Elmtree inlier (= Exploits), the lower part of the Tetagouche Group in the Mirimichi Highlands and possibly Cookson Formation (= Gander), and Avalon Zone. The boundaries between these major units appear to have originated as thrusts, however, in some cases, present boundaries are defined by transcurrent faults that truncate the nappe stack. From top to bottom, the original order of the nappes appears to have been: Notre Dame, Exploits, Gander, with the Burgeo thrust over the Exploits, although in places local reversals appear. The original thrust boundary is best preserved between the Exploits Subzone and Gander Zone. This boundary has been deformed by later upright to asymmetrical folds that together have produced a dome-and-basin interference pattern in south-central Newfoundland. The original vergence appears to have been generally towards the south. Some boundaries coincide with later strike-slip faults.

The earliest deformation is recorded in the obducted ophiolites as Early Ordovician. Description of the Ordovician - Early Devonian structures is subdivided by major nappe or zone: Notre Dame, Exploits, Gander, Avalon.

NOTRE DAME AND DASHWOOD SUBZONES (DUNNAGE ZONE)

The structure in the ophiolitic rocks of the Notre Dame and Dashwood subzones has not been studied in the same detail as some of those in the Humber Zone. The earliest recorded structures are discrete, amphibolite facies shear zones that vary in trend: NE-trending shear zones cut by NW-trending faults in the Annieopsquotch ophiolite complex and E-trending shear zones in the Deveraux Formation (Dunning and Krogh, 1985; Rast and Stringer, 1980; Flagler and Spray, 1988).

In the Twillingate area on the northern coast of Newfoundland, bimodal volcanic rocks of the Sleepy Cove Group are intruded by the Twillingate sodic granite (Williams and Payne, 1975). The granite is cut by a steeply dipping, NE- to E- and NW-trending, mylonitic foliation, which is most intense in the south but generally dies out towards the north. Williams and Payne (1975) correlate this foliation with a parallel foliation in the Sleepy Cove volcanic rocks, although Dewey *et al.* (1983) observed that it passes into the S_2 crenulation cleavage in the volcanic rocks. The mylonitic foliation in the granite is cut by undeformed mafic dykes.

The structure elsewhere in the eastern part of the Notre Dame and Dashwood subzones has only received detailed study in a few areas. For example, in Notre Dame Bay, Szybinski (1988, 1989) has documented the presence of several thrust duplex systems associated with folds and a penetrative cleavage in rocks of the Cutwell Group. The rocks were first affected by NE-verging thrusts associated with sinistral strike-slip faulting on the E-W Long Tickle Fault, followed by NW-vergent thrusting associated with sinistral oblique faulting. The E-trending Lobster Cove Fault truncates these structures, and kinematic indicators such as folds, shear bands and lineations indicate that it had a dextral, strike-slip movement (Szybinski, 1988; Calon and Szybinski, 1988). Traced westwards, the Lobster Cove Fault swings into a NE-trend and it becomes a SE-vergent thrust (Szybinski, 1989). Dean and Strong (1977) inferred a general southerly vergence for the Lobster Cove Fault based upon the regional geology.

In the Buchans area of central Newfoundland, a broad E-W culmination has provided an oblique cross-section across the structure (Calon and Green, 1987; Thurlow and Swanson, 1987; McClay, 1987). The structure appears to consists of a series of hinterland and foreland dipping, thrust duplexes that built up into a huge, E-trending, antiformal stack. The eyelid window outcrop pattern also indicates the presence of a complex system of frontal, oblique and lateral ramps for the thrust duplexes and a shingle-like stacking of the duplexes. The metamorphic grade decreases from amphibolite facies in the uppermost Hungry Mountain Complex through greenschist/epidote amphibolite facies in the next unit to non-metamorphic in all lower structural units. The downward decline in metamorphic grade was interpreted as the result of the hot emplacement of the Hungry Mountain Complex (Thurlow, 1981). A SSE direction of thrust vergence is indicated by the ramp geometry, orientation and vergence of minor folds, foliation and lineation attitudes and rare kinematic inducators such as S-C mylonites. Folding of structurally higher duplexes over the antiformal stack indicates that the structure formed as a piggy-back mechanism with deformation initiated by movement of the Hungry Mountain Complex. The Notre Dame and Dashwood subzones are separated by an E-W ductile shear zone (Little Grand Lake Fault). Kinematic indicators show that the Notre Dame subzone was thrust southwards over the Dashwood subzone (Van Berkel and Currie, 1988).

In southwestern Newfoundland, Dunning and Chorlton (1985) subdivided the rocks between the Long Range and Cape Ray faults into two ophiolitic nappes separated by an originally subhorizontal, amphibolite facies shear zone. Thrusting predates intrusion of undated gabbro, diorite and granodiorite.

NOTRE DAME/DASHWOOD - EXPLOITS BOUNDARY

In Notre Dame Bay, the Notre Dame/Dashwood - Exploits boundary is defined as the Lukes Arm - Sops Head Fault (Williams *et al.*, 1988). This fault follows a narrow zone of melange (Boones Point Complex) that separates Lower Ordovician volcanic and volcaniclastic rocks to the north from Upper Ordovician turbidites to the south (Nelson, 1981). The fault cuts across two earlier fold phases observed in the surrounding area: isoclinal, recumbent folds refolded by tight asymmetric folds associated with a transecting cleavage both inferred to have developed under an overall sinistral shear regime (Karlsrom *et al.*, 1982; Blewett, 1989). The tight folds and associated crenulation cleavage followed by open NW-trending folds and kink bands are preferentially developed in the fault zone and are inferred to have been produced in an E-W dextral shear zone. This fault continues westwards as the Tommy's Arm - Crescent Lake Fault which have similar dextral shear structures (Blewett and Pickering, 1988).

The Notre Dame/Dashwood - Exploits boundary may be traced inland to the southwest as a poorly exposed, steep brittle fault characterized by mylonites and catalasis (Brown, 1977; Swinden and Sacks, 1986; Williams *et al.*, 1988). Kinematic studies of the Cape Ray Fault suggest early NW-vergent thrusting followed by westward oblique thrusting (Williams *et al.*, 1989).

In the Elmtree inlier of northern New Brunswick, the boundary between the Deveraux and Pointe Verte Formations is marked by mylonitic amphibolite that is locally shallowly dipping with kinematic indicators showing a SE- or S-directed thrusting (Van Staal *et al.*, 1988a; Langton and Van Staal, 1989). The Melange Formation, overlying the Devereaux Ophiolite Complex in the Elmtree inlier of northern New Brunswick, possesses a melange foliation deformed about NE-SW folds associated with an axial planar cleavage (Rast and Stringer, 1980).

EXPLOITS SUBZONE

Ophiolitic complexes occur in several locations along the boundary between the eastern Exploits Subzone and the Gander Zone. The nature of the structures within them is largely undescribed. Ophiolitic melange units in the Exploits Subzone are all Ordovician in age. They generally possess one or more scaly cleavage and intrafolial folds ± refolded folds and occur at several locations within the southeastern Dunnage Zone, e.g. Dunnage, Carmanville and Cold Spring Melanges. Although the nature of the structures within them has generally not been studied in detail, they are important to the structural history because they record nappe emplacement and may separate different nappes. The Dunnage Melange in northern Newfoundland consists of sedimentary and igneous blocks in a shaly matrix (Hibbard and Williams, 1979; Kay, 1976). The matrix displays wispy folds and a scaly cleavage which are inferred to have been produced during deposition. The ophiolitic Carmanville Melange occurring in the Davidsville Group between the GRUB line and Dunnage Island has been correlated with the Dunnage Melange (Pajari *et al.*, 1979; Karlstrom, 1982). The Cold Spring melange (dunite olistoliths in a chaotic matrix) occurs along the Exploits - Gander boundary in southern Newfoundland (Piasecki *et al.*, 1990). The melange horizons have generally been assumed to be olistostromes (Pajari *et al.*, 1979), however, Karlstrom (1982) interprets them as thrust zones between nappes. This is borne out on New World Island where several large thrust zones are coincident with non-ophiolitic olistostromes, e.g. the Cheneyville Melange, Carters Cove olistostrome, Joeys Cove Melange, Boones Point Complex and Toogood Formation (Nelson, 1981; Van der Pluijm, 1984; Karlstrom *et al.*, 1982). Associated minor structures include tight-isoclinal intrafolial folds and refolded folds ± an associated cleavage or cleavages and thrusts. NW-trending sheath folds and the orientation of horses in a thrust duplex on New World Island have been interpreted in terms of both SE- and NW-directed thrusting (Van der Pluijm and Williams, 1985; van der Pluijm, 1986). The recumbent structures in New World Island are deformed by (1) upright, NE-SW folds associated with a regional cleavage; (2) ductile, dextral strike-slip, NE-SW faulting associated with asymmetric folds and a foliation that are overprinted by the Loon Bay Pluton; and (3) brittle, sinistral, NE-SW faulting that deform dacitic dykes correlated with the Loon Bay Pluton (Kay, 1976; Van der Pluijm, 1986; Williams *et al.*, 1988).

In central Newfoundland, the Victoria Lake and Baie d'Espoir Groups, have been deformed into open-isoclinal, generally NE-trending subhorizontal folds with an axial planar cleavage formed under greenschist facies (- local lower amphibolite facies) metamorphic conditions, that have been refolded by upright folds accompanied by a crenulation cleavage (Colman-Sadd, 1980; Blackwood, 1985). The basal contact of the Victoria Lake Group is inferred to be a thrust with Late Precambrian rocks in the footwall (Evans *et al.*, 1990). Similar, early recumbent folds and thrusts also occur at the contact with the Gander Zone.

In southwestern Newfoundland, east of the Cape Ray fault zone, Chorlton (1978, 1983), O'Brien (1988) and O'Brien *et al.* (1988) reported rare early recumbent folds associated with an axial planar foliation and NW-vergent thrusts (now steeply dipping) developed under amphibolite facies metamorphic conditions in the Bay du Nord Group. Stretching lineations plunge steeply southeastwards.

These early structures affect the La Poile pluton and early phases of the North Bay granite suite, and were overprinted by oblique sinistral, NE-SW to E-W shear zones and associated structures (steeply SE-dipping schistosity-crenulation cleavage, variably plunging folds and gently NE-dipping extension lineation) developed under amphibolite facies metamorphic conditions. These structures post-date contact metamorphic minerals in the aureole of the North Bay pluton, although some biotite grew along the crenulation cleavage. The latest phases of the North Bay granite suite post-tectonically cut across all the structures in the country rocks.

In northern New Brunswick, rocks of the Elmtree inlier have been correlated with those in the northern Mirimichi Highlands (Van Staal, 1987, 1988; Van Staal *et al.*, 1988a). The major structure of the Elmtree inlier consists of three nappes (Langton and van Staal, 1989): (1) the upper nappe (correlated with the Notre Dame subzone) consists of the Deveraux Formation; (2) the central nappe (correlated with the Exploits subzone) comprises the late Arenig - early Caradocian, Pointe Verte Formation (alkali basalt, slate and wacke); and (3) the lowest nappe (correlated with the Exploits sunzone) is made up of wacke, shale, and minor basalt and ultramafic rocks (late Ordovician Elmtree Formation). Based upon contrasting degrees of deformation, Rast and Stringer (1980) inferred the presence of an angular unconformably between these Ordovician rocks and the Llandoverian C_3-C_4 Armstrong Brook conglomerates, however, the exposed contact is tectonic (Noble, 1986). On the other hand, van Staal (1987; 1988a) has documented that the Ordovician and Silurian rocks have a similar structural history. The few kinematic indicators in gently dipping shear zones in the Fournier Complex suggest southerly to southeasterly direction of obduction (van Staal, 1987; Langton and van Staal, 1989).

The major structure in the northern Mirimichi Highlands consists of a series of polydeformed thrust slices from top to bottom: (1) undated wackes that maybe equivalent to the Pointe Verte Formation; (2) Caradocian graphitic shale, siltstone and chert; (3) late Caradocian graphitic shale, siltstone and chert overlying early-middle Caradocian red-green phyllite interbedded with ocean floor basalt; and (4) late Arenig - early Llanvirn phyllite and quartzite correlated with the Gander Zone (van Staal, 1988b). The Tetagouche Group has suffered polyphase deformation which is generally more complex and at higher metamorphic grades in the northeast than in the southwest (Helmstaedt, 1970; Fyffe, 1982; Venugopal, 1978, 1979, 1981, 1982; Crouse, 1977, 1979, 1981a and b; Irrinki, 1978, 1979, 1980, 1981; St. Peters, 1980, 1982; Lutes, 1979, 1981; van Staal and Williams, 1984; van Staal; 1987; van Staal *et al.*, 1988). The earliest structures consists of intrafolial isoclinal sheath folds, micaceous bedding-parallel foliation, stretching lineations and thrusts which were refolded by variably oriented, tight-isoclinal folds with an axial planar slaty cleavage or crenulation cleavage and thrusts. These structures are sequentially overprinted by a differentiated crenulation cleavage, a fracture-crenulation cleavage and sporadic recumbent kinks. Unfortunately, the orientation of the sheath folds was not reported, however, van Staal (1987) argued for easterly to southeasterly thrusting based upon regional considerations.

The minor structure in the Elmtree inlier and the northern Mirimichi Highlands consists of a generally composite foliation, a stretching lineation and isoclinal folds overprinted by NE- or E-trending crenulation cleavage associated with tight steeply plunging folds followed by a spaced cleavage axial planar to open kinks and folds (Van Staal *et al.*, 1988a; Langton and Van Staal, 1989).

EXPLOITS-GANDER BOUNDARY

The Exploits-Gander boundary is generally a subhorizontal thrust boundary folded by upright to steeply dipping folds of different orientations which produce a dome-and-basin outcrop pattern. Thus, the main Noel Paul's - Grub Fault Line is sinuous and reappears around domes such as Mount Cormack. The immediate hanging wall of the thrust is commonly marked by disrupted ophiolitic remnants that display tectonic fabrics which have received little detailed structural study (Kennedy, 1975, 1976; Blackwood, 1979; Pajari *et al.*, 1979; Dewey *et al.*, 1983; Colman-Sadd and Swinden, 1984; Williams *et al.*, 1989). The initial sense of movement on the Exploits-Gander boundary varies depending upon its orientation. Thus, E-trending boundaries are generally subhorizontal to gently dipping and kinematic indicators show dip-slip movements (e.g. around Dollard Pond: Piasecki *et al.*, 1990). As the boundary swings towards a NE-trend, its dip steepens and the sense of shear becomes sinistral transcurrent (e.g. in northeastern

Newfoundland, the GRUB line is associated with a 1 km wide zone of mylonites and ultramylonites developed from both Davidsville and Gander Groups with S-C fabrics and shear bands that indicate south to southwest sinistral shear along a strike-slip thrust (Hanmer, 1981; Piasecki, 1988); in southern Newfoundland, the Day Cove Thrust shows sinistral shear fabrics in a gently dipping strike-slip thrust (Colman-Sadd, 1976; Piasecki, 1988; Piasecki *et al.*, 1990)). On the other hand, where the boundary swings northwestwards, it tends to steepen and the movement becomes dextral transcurrent (e.g. in south-central Newfoundland, the Great Burnt Lake Fault defines a S-trending part of the boundary, and subhorizontal lineations in a steep planar fabric with kinematic indicators such as S-C foliations, shear bands, feldspar porphyroclasts with tails, that imply dextral transcurrent movements followed by dip-slip movements with downthrow to the east) (Colman-Sadd, 1985; Piasecki *et al.*, 1990). Another example of such a transition occurs in Bay D'Espoir shear zone that remains steeply dipping throughout as it swings from NE- through W- to WNW-trends as traced from east to west, and the sense of shear becomes progressively reversed from dominantly sinistral through both sinistral and dextral to mainly dextral, respectively (Piasecki *et al.*, 1990). The Bay d'Espoir shear zone is syn-kinematically intruded by the Gaultois granite (Blackwood, 1985). The Exploits-Gander contact may be affected by later reverse faulting that places the Gander Zone above the Exploits Subzone, e.g. Noel's Paul Brook and Coy Pond (Colman-Sadd and Swinden, 1984; Colman-Sadd, 1987). These later NW-vergent thrusts also affect the Rogerson Lake Conglomerate.

The Exploits - Gander boundary in the northern Mirimichi Highlands is the Bathurst Fault (Van Staal *et al.*, 1988). It is marked by a narrow phyllonite zone with lenses of glaucophane schist, across which there are opposite facing directions. Reliable kinematic indicators are lacking, partly due to overprinting by later upright structures, however the presence of glaucophane schist on top of lower pressure greenschist facies rocks suggests it is a major thrust fault with an E- or SE-vergence based upon regional arguments (Van Staal, 1987; Van Staal *et al.*, 1988). Structural studies on this boundary elsewhere in the Mirimichi Highlands have not yet been carried out.

GANDER ZONE

Structural observations along the eastern Gander Zone in Newfoundland show that it is a NE-trending sinistral ductile shear zone varying in width from 60 kms in the north to 12 kms in the south (Hanmer, 1981; Piasecki, 1988). The ductile shear zone changes from gently inclined in the west to steeply dipping adjacent to the Dover-Hermitage Bay Fault Zone. This steepening is accompanied by a regional increase in metamorphic grade from greenschist to upper amphibolite facies and migmatization. Mylonites in the shear zone contain abundant kinematic indicators such as C-S fabrics and shear bands, noncylindrical folds upto sheath folds with sheath axes parallel to the regionally penetrative stretching lineation defined by mineral alignment, boudin long axes, quartz pressure shadows on pyrite grains, and rotated porphyroblasts or porphyroclasts. Various observations indicate that the ductile shear deformation was synchronous with emplacement and cooling of granitoid plutons: (1) the angle between the granite sheets and the foliation generally varies with the intensity of the internal foliation in the sheets and all permutations were observed; (2) the foliation in the granites is defined by the alignment of polycrystalline quartz leaves, biotite aggregates, and feldspar megacrysts or aggregates derived from them, however some K-feldspar megacrysts are decussate and overgrow post-foliation shears and the margins of the granite sheets; and (3) concentric and triangular foliation trajectories produced by diapiric plutonism.

Rocks correlated with the Gander Zone in the Mount Cormack window in central Newfoundland are cut by a vertical cleavage within which tight-isoclinal folds axes vary from horizontal to vertical (dominant), locally refolded by a crenulation cleavage (Colman-Sadd and Swinden, 1984). The low pressure metamorphism increases towards the centre of the window from greenschist to upper amphibolite facies with local migmatization and the metamorphic isograds cut across the major structure. The migmatitic gneisses in the centre of the window display a more complex structural geometry with refolded folds. The dominant structure is cut by several small granite plutons.

Gander Zone rocks in the Meelpaeg region of south-central Newfoundland are also affected by complex folds and metamorphism that ranges from greenschist near the Gander-Exploits boundary to upper amphibolite facies (Colman-Sadd, 1985, 1987). The structures were interpreted in terms of polyphase deformation rather than in terms of shear zone fabrics thereby making kinematic interpretations suspect.

In the Miramichi Highlands of New Brunswick, the lowest member of the Tetagouche Group, the Nepisiquit Formation and the quartzite-slate/phyllite unit, are generally correlated with the Gander Zone. Although its contact with the overlying Llanvirnian- middle Arenigian unit of quartzite pebble conglomerate, calcareous siltstone and tuff is generally concordant, in the southern Miramichi Highlands, Irrinki (1980) reported an unconformable contact where mesoscopic folds in the quartzite-slate unit are absent in the overlying unit. The earliest structure in the quartzite-slate unit consists of a slaty cleavage in the south becoming a micaceous schistosity in the central Miramichi Highlands. It appears to be absent in all the succeeding units. Its present geometry is largely due to subsequent deformation and initial kinematic indicators are presently unknown. In the northern Miramichi Highlands, the Nepisiquit Formation suffered the polyphase deformation common to other Late Ordovician rocks (Van Staal *et al.*, 1988).

In southern New Brunswick and adjacent parts of Maine, the Cookson Formation has been deformed by early isoclinal folds of variable trend associated with an axial planar micaceous cleavage refolded by reclined, west-vergent folds accompanied by a spaced cleavage and thrusts (Ludman, 1985). These structures were not recorded in the unconformably overlying Siluro-Devonian rocks around Oak Bay.

GANDER/EXPLOITS-AVALON BOUNDARY

In Newfoundland, the Gander-Avalon boundary is placed at the Dover-Hermitage Bay Fault and has generally been extrapolated southwestwards into the Gulf of St Lawrence. However, the discovery of rocks with Avalonian affinities along the south-central coast of Newfoundland (Burgeo Terrane) indicates that the western boundary of the Avalon reappears as the Dragon Bay-Bay d'Est fault zone (O'Brien *et al.*, 1986, 1989; Keppie and Dallmeyer, 1989). Here the Gander Zone rocks are absent and the Exploits subzone is juxtaposed against the Avalon Zone (Burgeo terrane). In central Newfoundland, two late Proterozoic plutons lie in windows structurally beneath rocks of the Exploits subzone (Evans *et al.*, 1990). The nearest correlatives of these plutons occur in the Burgeo terrane.

The Dover Fault is <400 m wide zone across which there is a marked contrast in the intensity of deformation from intense polyphase deformation accompanied by upper amphibolite facies metamorphism in the Gander Zone to milder greenschist facies deformation in the Avalon Zone (Dallmeyer *et al.*, 1981). Away from the fault zone, kinematic indicators in both the Gander and Avalon Zones have been interpreted in terms of sinistral transpression at higher metamorphic grades (Hanmer, 1981; Caron and Williams, 1988a). On the other hand, microstructures in the ENE-trending part of the fault zone indicate dextral shear followed by normal, reverse and dextral displacements taking place under metamorphic conditions gradually falling from lower greenschist facies to brittle conditions (Hanmer, 1981; Caron and Williams, 1988b). Hanmer (1981) interpreted the Dover Fault as a dextral R' reidel shear within the regional, NNE-trending, sinistral shear zone. S-C fabric and shear bands associated with the NE-trending Hermitage Bay Fault also indicate sinistral movements (Piasecki, 1988).

The Dragon Bay and Bay d'Est faults are vertical and swing from WNW along Dragon Bay to WSW along Bay d'Est (Blackwood, 1984, 1985; O'Brien *et al.*, 1986). The Dragon Bay Fault cuts across granitoid rocks as a ca. 1 km wide mylonitic-ultramylonitic zone. The associated foliation in the granitoid rocks outside the fault zone trends anticlockwise relative to the fault and swings into the fault zone suggesting a dextral component of movement (Piasecki *et al.* 1990). Blackwood (1984, 1985) inferred that the Dragon Bay Fault cut across previously recumbent south-vergent folds that placed the Bay du Nord Group (Dunnage Zone) over the Burgeo Terrane (Avalon Zone). The Bay d'Est Fault varies from vertical to SE-dipping and is associated with variably plunging folds cut by brittle fault breccia (O'Brien, 1989). The fault juxtaposes Silurian La Poile Group on the south against Ordovician Bay du Nord Group on the north, and O'Brien (1989) inferred a N-vergence for the early thrusting. Relationships between the structural fabrics and various granites indicate that plutonism overlapped the deformation event. The Burgeo granitoid pluton intrudes both the Bay du Nord Group (Exploits subzone) and the La Poile Group (Avalon Terrane) and so may be considered a stitching pluton, however, it is generally strongly sheared and is inferred to be syntectonic (O'Brien *et* *al.*, 1986). Porphyroblasts in the aureole of the North Bay granite overprint S_1 and S_2 foliations, but dykes of the granite are deformed by F_2 folds indicating that it was late syn-D_2.

In southern New Brunswick, the Basswood Ridge - Cox Brook Fault marks the northern limit of Silurian-Gedinnian rocks containing Rhenish fauna and so defines the northwestern margin of the Avalon Zone in southern New Brunswick (Fyffe, 1989; Keppie *et al.* 1991). Although no kinematic analysis of the fault zone has been undertaken, the associated asymmetric folds led Fyffe (1989) to suggest that it is a SE-vergent thrust, however, parallel faults within the Avalon Zone have suffered both sinistral Silurian and dextral Devonian displacements (Leger and Williams, 1986). The fault zone has been post-tectonically intruded by late Devonian porphyry dykes. The continuation of this fault in Maine has been stitched by several plutons, the oldest of which is the Lucerne pluton (Loiselle *et al.*, 1983).

AVALON ZONE

South of the Bay d'Est-Dragon Bay fault zone in southern Newfoundland (Burgeo Terrane), late Proterozoic-Ordovician rocks with Avalonian affinities are unconformably overlain by the 3-5 km thick, Silurian La Poile Group inferred to have been deposited in half-graben above SE-dipping listric normal faults (O'Brien *et al.*, 1986; 1989; Dunning and O'Brien, 1989). Deformation of the La Poile Group took place under greenschist facies metamorphic conditions and varies from a weakly developed fracture cleavage associated with open folds in the east to polyphase deformation involving coaxial folds associated with a slaty cleavage overprinted by a crenulation cleavage followed by cross folds (Chorlton, 1980; O'Brien, 1988). The major structures are upright folds (locally overturned to the northwest) and N-vergent listric thrusts (reactivated normal faults). The composite Burgeo Intrusive Suite contains both syn-D_1 foliated phases and post-D_1 weakly foliated phases, whereas the Chetwynd and Francois granites are post-tectonic (Dunning *et al.*, 1990).

The rocks of the Avalon Zone adjacent to the Dover-Hermitage Fault in Newfoundland, display a low grade cleavage that dies out eastwards. This deformation appears to be genetically connected to the folding and SE-thrusting observed elsewhere in the Newfoundland Avalon Zone. The Upper Devonian Great Bay de l'Eau Formation appears to be relatively undeformed, although it was also affected by some of the thrusting (Williams, 1971; Greene, 1975; O'Driscoll and Strong, 1979).

In the Cape Breton Highlands, the Money Point Group, Jumping Brook complex and Sarach Brook metamorphic suite (Macdonald and Smith, 1980;Currie *et al.*, 1982; Dunning *et al.*, 1990) may correlate with the La Poile Group of southern Newfoundland. A structural cross-section across the Cape Breton Highlands reveals a N-trending, vertically dipping, high metamorphic grade, central zone bounded by outward verging asymmetric to recumbent, low metamorphic grade, marginal zones. This is interpreted as a positive flower structure (Keppie *et al.*, in press). The eastern margin of this flower structure passes though the Money Point area. Here, the Money Point and Cape North groups have been deformed by: (D_1) subhorizontal, isoclinal folding; regional phyllitic to gneissic foliation and penetrative lineation under prograde, greenschist facies metamorphism; (D_2) subhorizontal, mesoscopic folds; moderately-steeply dipping, penetrative crenulation foliation and subhorizontal lineation; immediately followed by peak, middle greenschist (east) - upper amphibolite (west) facies metamorphic conditions with mapped isograds subparallel to lithological contacts (Macdonald and Smith, 1980). Lineations L_1 and L_2 are parallel to one another and to the long axes of stretched clasts in meta-conglomerate and meta-tuff with X:Y:Z ratios of 12:2:1. In the Cape North Group and the Cape North pluton the S_1 and S_2 fabrics appear to be spatially related to ductile shear zones, and may be interpreted as C-S fabrics they indicate a sinistral sense of shear (Keppie *et al.*, in press).

The western side of the positive flower structure occurs in the Cheticamp area. Here, the Jumping Brook complex has suffered varying degrees of penecontemporaneous deformation (Craw, 1984). The major structure consists of a steeply dipping high grade rocks in the east that were thrust up and westwards over lower grade generally flat-lying rocks to the west. Polyphase structures in the east with down-dip stretching lineations are give way to single phase, W-vergent, recumbent, gently N-plunging, isoclinal folds in the west.

In the south-central part of the Cape Breton Highlands, the N-trending, sub-vertical Southern

Highlands shear zone displays stretching lineations varying from vertical to horizontal with downthrows to the west (Mengel *et al.*, 1991). This contrasts with the Eastern Highlands shear zone that displays oblique dextral kinematic indicators with a downthrow to the southeast (Lin and Williams, 1990; Mengel *et al.*, 1991). Such variations in movement direction and throw are typical of transcurrent shear zones, and confirm the general positive flower structure deduced for the Cape Breton Highlands (Keppie *et al.*, in press).

Structures in the Cambro-Ordovician rocks in the Avalon Zone are generally simple, however, there are local zones of intense deformation adjacent to faults. Age constraints on these structures are usually imprecise and an Ordovician, Silurian or Devonian age is possible. However, in the Antigonish Highlands, an Ordovician age can be documented (Keppie and Murphy, 1988). Here, a local zone of complex polyphase deformation occurs in an area where two en echelon NE-SW vertical faults overlap. Structures in the overlap area consist of thrusts and open-isoclinal, recumbent, NW-trending folds with a micaceous cleavage refolded by open-tight, upright-asymmetric, NW-SE folds. These structures die out rapidly in all directions and they are unconformably overlain by early Llandoverian rocks of the Arisaig Group. The structures are inferred to have been produced by sinistral motions on the bounding NE-trending faults. This movement expelled the Cambro-Ordovician rocks from a pull-apart basin by first thrusting them northeast and then back-thrusting them.

In southern New Brunswick, the ductile dip-slip movements along the Belleisle Fault produced polyphase structures in rocks as young as Gedinnian but do not affect the early Carboniferous Beaver Head Group (Brown and Helmstaedt, 1970). However, Garnett and Brown (1973) showed that the Belleisle ductile shear zone grade from undeformed, through an intermediate zone of horizontal stretching into a highly strained zone with vertical stretching. Although they interpreted this in terms of a single protracted period of heterogeneous strain, a model of horizontal shear superimposed by vertical shear seems more likely. Just south of the Bellisle Fault is an associated shear zone: the Pocologan mylonite zone. Although there has been much debate about the age and movement history of its zone (e.g. Rast and Dickson, 1982; Leger and Williams, 1986), recent geochronological data indicates that the associated Kinston dyke complex is Early Silurian: 435.5 ± 1.5 Ma U-Pb zircon age (Doig *et al.*, 1990). Based on the orientation of the dykes anticlockwise of the NE-trending shear zone, Doig *et al.* (1990) have inferred that the dykes were emplaced in a sinistral shear regime. S-thrusting and folding in the adjacent Avalon Zone may be of the same age (Ruitenberg and McCutcheon, 1982; McLeod and McCutcheon, 1981).

AGE OF THE STRUCTURES

Penoscotian Episode of Deformation: Early Ordovician Deformation Related to the Accretion of the Exploits Subzone and the Gander Zone

The oldest structures recorded in the Exploits Subzone are Early Ordovician and they appear to record the obduction of Exploits ophiolite upon the Gander Zone and a late Proterozoic continental unit (Valentine Lake and Crippleback Lake plutons). A lower age limit on the time of obduction is provided by the 494 ± 3 Ma U-Pb zircon age on trondjhemite in the allochthonous Pipestone Pond Ophiolitic Complex (Dunning and Krogh, 1985). The nearby Cold Spring melange which contains olistoliths of dunite, quartzite and metapelite in a phyllonitic matrix is overlain by the late Arenig-early Caradocian Bay d'Espoir Group that contains a few ophiolitic trondjhemite and gabbro clasts (Colman-Sadd and Swinden, 1984). This brackets the time of obduction to the late Tremadocian and early Arenigian. A similar time of deformation is recorded in the Dunnage Melange whose matrix contains Tremadocian graptolites (Hibbard *et al.*, 1977). The Middle Cambrian and Arenigian conodonts recovered from blocks in different parts of the melange (Kay and Etheridge, 1968; Hibbard *et al.*, 1977) indicate that some parts of the melange must be as young as Arenigian. An upper limit on the scaly cleavage in the melange is provided by the 445 ± 8 Ma Rb-Sr whole rock isochron on the cross-cutting Coaker porphyry (Lorenz, 1984), and the 434 to 441 ± 13 Ma K-Ar biotite ages from the Causeway diorite (recalculated from Kay, 1976). An upper limit on the time of obduction of the Exploits ophiolite along the Gander River Ultrabasic Belt (GRUB) is given by the age of the oldest, unconformably overlying rocks. At Gander Lake, conglomerates just above the unconformity contain pebbles of schistose serpentinite, gabbro, diabase, amphibolite, phyllite and quartzofeldspathic schist, and detrital chromite, garnet and mica (McGonigal, 1973; Kennedy, 1976; Dewey *et al.*, 1983). Overlying sediments have yielded brachiopods and a trilobite of possible late Arenigian age (Neuman, 1984; McKerrow and Cocks, 1977; Jenness, 1963). Further north at Weir's Pond, a limestone yielding late Llanvirnian-early Llandeilian conodonts (Stouge, 1980) lies unconformably upon serpentinite (Blackwood, 1978). Associated sediments include conglomerates containing pebbles of ultramafic, gabbroic, basaltic, trondjhemitic and agglomeratic rocks. At Mount Cormack, chromite, serpentinized peridotite and possible trace volcanogenic detritus presumably derived from allochthonous Exploits ophiolite (Coy pond and Great Bend) occur in late Llanvirn to early Llandeilo limestone conglomerate that (?)unconformably overlies the Spruce Brook Formation (Dec and Colman-Sadd, 1990).

To date, complimentary Early Ordovician deformation in the Gander Zone in Newfoundland has not been recorded, probably due to intense subsequent deformation. However, early Ordovician deformation is recorded in the south-central Mirimichi Highlands. Here, Tremadocian black shales occur at the top of the quartzite-slate unit of the Tetagouche Group (Fyffe *et al.*, 1983), and it is possible that they record the sudden deepening of the basin related to obduction of the Exploits subzone upon the Gander Zone. An early cleavage in the quartzite-slate unit of the Tetagouche Group appears to be absent in all the succeeding units. A lower limit on the age of cleavage is given by the age of quartzite-slate unit which contains late Tremadocian graptolites (Fyffe *et al.*, 1983) and *Oldhamia* of probable Cambrian age (Neuman, 1968). While its contact with the overlying unit of quartzite pebble conglomerate, calcareous siltstone and tuff containing Llanvirnian - middle Arenigian brachiopods (Neuman, 1968; Poole, 1963) is generally concordant, Irrinki (1980) reported an unconformable contact where mesoscopic folds in the quartzite-slate unit are absent in the overlying unit: this provides an upper limit on the time of deformation. The conglomerate-siltstone-tuff unit is overlain, locally unconformably, by a slate (± bimodal volcanic rock) unit (= Belle Lake Slate) containing a late Llanvirnian to early Caradocian fauna (Venugopal, 1979; Fyffe *et al.*, 1983). The slate unit is conformably overlain by bimodal volcanic rocks or an undated greywacke unit containing phyllite clasts with a pre-depositional fabric presumably derived from the quartzite-slate unit at the base of the Tetagouche Group (Crouse; 1981a, 1981b; Irrinki, 1980, 1981). This early Arenigian deformation has been called the Penobscotian Disturbance/Orogeny in the nearby parts of Maine (Neuman, 1968; Hall, 1969). An upper limit on the earliest foliation in the central Miramichi Highlands is also provided by the presence of sedimentary xenoliths containing a foliation that predates a foliation in the host South Renous River Granite (Irrinki, 1980). This granite is intrusive into the quartzite-slate unit which appears to be the source of the xenoliths and has yielded a U-Pb zircon age of 448 +17/-16 Ma (Bevier, 1988). Just north of the South Renous River Granite occur several similar deformed granites also intrusive into the quartzite-slate unit. These deformed granites have yielded a Rb-Sr whole rock isochron of 479 ± 14 Ma (Fyffe *et al.*, 1977) and U-Pb zircon ages of 452 +15/-1 Ma (Fox Ridge pluton), 455 +12/-9 Ma (Serpentine River pluton), 458 +16/-1 Ma (Mullin Stream pluton), and 470 ± 1 Ma (Meridian Brook pluton)(Bevier, 1988) which places an upper limit on the pre-existing foliation in the xenoliths. Whalen (1987) inferred that these plutons were consanguinous with the felsic volcanic rocks in the Tetagouche Group.

In the Notre Dame and Dahwood Subzones, similar Tremadocian to early Arenigian time of deformation are documented in the Twillingate area on the northern coast of Newfoundland and in the Annieopsquotch ophiolite complex. At Twillingate, variably trending, steeply dipping mylonitic foliations in the Twillingate granite, dated at 510 + 17 Ma (U-Pb zircon upper intercept age: Williams *et al.*, 1976), is cut by undeformed mafic dykes yielding an ^{40}Ar/^{39}Ar plateau age on amphibole of 482 ± 9 Ma (recalculated from Williams *et al.*, 1976).

In the Annieopsquotch ophiolite complex, trondjhemite intruded along NW-SE faults that cut NE-trending amphibolite facies shear zones, have yielded U-Pb zircon ages of 481 +4/-2 Ma and 478 +3/-2 Ma on trondjhemite (Dunning and Krogh, 1985). Thus, development of these structures started prior to 481 +4/-2 Ma in an intra-oceanic environment.

LATE ORDOVICIAN - EARLY DEVONIAN STRUCTURES

Deformation during Late Ordovician, Silurian and Early Devonian times may be related to two major accretionary events: obduction of the Notre Dame-Dashwood over the Exploits subzones (Beothukan), and the accretion of the Avalon Zone with the rest of the central mobile belt (Humberian).

Beothukan Episode of Deformation: Late Ordovician - Early Devonian Generally SE-Vergent Structures Associated with Amalgamation Between the Notre Dame/Dashwood and Exploits Subzones

A Middle Ordovician (Llanvirn-Llandeilo) lower limit on the time of earliest accretionary deformation in the Notre Dame and Dashwood subzones is given by the age of units involved in the deformation: (1) the late Arenig-early Llanvirn conodonts and graptolites in the Catchers Pond and Cutwell Groups (O'Brien and Szybinski, 1989; Williams, 1989); (2) the 473 ± 2 Ma U-Pb zircon age for the Roberts Arm Group (Dunning *et al.*, 1987); (3) the 479 ± 3 Ma U-Pb zircon age of the Mansfield Cove Complex (Dunning *et al.*, 1987); (4) the 473 +3/-2 Ma U-Pb age of the Buchans Group that also contains late Arenig - early Llanvirn conodonts (Nowlan and Thurlow, 1984; Dunning *et al.*, 1987); (5) the 478 +3/-2 and 481 +4/-2 Ma U-Pb zircon ages of the Annieopsquotch ophiolite complex (Dunning and Krogh, 1985); and (6) the 461 ± 3 and 463 +3/-2 Ma U-Pb zircon ages of the ophiolitic rocks of the Deveraux Formation exposed in the Elmtree inlier of northern New Brunswick (van Staal *et al.*, 1988; Flagler *et al.*, 1989). In the immediately adjacent Exploits subzone, a similar Llandeilo- Caradoc lower limit on the time of earliest accretionary deformation is given by the age of units involved in the same deformational events: (1) Cobbs Arm Limestone (Llandeilo-Llanvirn) and Suumerford Group (Tremadoc-Llandeilo) (Bergstrom *et al.*, 1974; Fahraeus and Hunter, 1981; Arnott *et al.*, 1985); (2) the 513 ± 2, 498 +6/-4, and 462 +4/-2 Ma U-Pb zircon ages recorded for various parts of the Victoria Lake Group, the latter associated with Llanvirn-early Llandeilan conodonts (Kean and Jayasinghe, 1982; Evans *et al.*, 1990); (3) the late Arenig to early Caradocian graptolites and conodonts in the Pointe Verte Formation (Fyffe, 1986; Nowlan in van Staal *et al.*, 1988; Langton and van Staal, 1989); and (4) early-late Caradocian graptolites and conodonts in the Tetagouche Group (Kennedy *et al.*, 1979; Fyffe *et al.*, 1983; Skinner, 1974; van Staal *et al.*, 1988).

Melanges and olistostromes are intimately related to thrust zones and probably formed as debris flows and olistostromes at the leading edge of the thrust sheets which may become the locus of thrust zones as the allochthons progressively advance. In this case, the matrix of the olistostromal melanges dates the time of thrusting, while the youngest clast provides a lower limit on the time of movement. The Boones Point Complex is distributed along the Lukes Arm - Sops Head Fault zone, the Notre Dame - Exploits subzone boundary (Nelson, 1981; Blewett, 1989). Several associated thrust zones are coincident with olistostromes, e.g. the Cheneyville Melange, Carters Cove olistostrome (CCI), Joeys Cove Melange and Toogood Formation (Van der Pluijm, 1984; Karlstrom *et al.*, 1982). These non-ophiolitic melange/olistostromes generally consist of blocks of basalt, tuff, conglomerate, greywacke, sandstone, chert, argillite, limestone and black slate in an argillaceous matrix. In addition, the Boones Point Complex also contains clasts of rhyolite, intermediate and silicic plutonic rocks inferred to have been derived from the Roberts Arm Group (Nelson, 1981). The matrix of these olistostromal melanges ranges from early Caradocian in the Carters Cove olistostrome, through Ashgillian in the Intricate Harbour olistostrome to early late Llandovery in the Joeys Cove Melange (McKerrow and Cocks, 1981; Arnott, 1983). The age of the clasts in the Carters Cove olistostrome is probably Tremadocian (Horne, 1970), whereas in the Intricate Harbour olistostrome they are late Caradocian: D. clingani Zone (Bergstrom *et al.*, 1974). The age of the Boones Point Complex also falls within Caradocian- Llandoverian interval because it contains limestone clasts with late Llanvirnian-Llandeilian conodonts (Stouge, 1980) and it appears to conformably overlie a greywacke-slate unit (Point Leamington and Gull Island Formations) containing middle Caradocian - early Ashgillian graptolites (Bergstrom *et al.*, 1974). Debris flow horizons interbedded within this unit are identical to the Boones Point Complex (Nelson, 1981) and contain limestone blocks with corals of probable Ashgillian age (Arnott *et al.*, 1985). These data suggest an Ashgillian age for the Boones Point Complex.

Another stratigraphic feature that is synchronous with the initial emplacement of thrust sheets is the sudden deepening of the basin in front of the nappes produced in response to loading of the lithosphere (Quinlan and Beaumont, 1984). This may be recorded in the stratigraphic record by black shales overlain by greywackes in the basin. In the Exploits subzone in Newfoundland and New Brunswick, black shales first appear in the Nemagraptus gracilis zone (earliest Caradocian) with greywackes first appearing in the D. multidens, D. clingani and P. linearis zones (Erdtmann, 1976; Arnott *et al*., 1985; van Staal, 1988b; Williams, 1989). Thus, the Notre Dame nappes began to be emplaced in earliest Caradocian times. Olistostromal melanges indicate that it continued through late Ordovician and Early Silurian times. However, that it continued until at least Ludlovian times is documented by the presence of similar thrust zones and associated intrafolial folds in rocks as young as the Llandoverian - Wenlockian Indian Islands Group and the early Ludlovian - Llandoverian Botwood Group (Karlstrom *et al*., 1982; Dean, 1978), and in early Silurian rocks in northern New Brunswick (van Staal, 1987, 1988a). The structures in northern Newfoundland are cut by granitoid plutons dated at 393 ± 30 Ma and 383 ± 15 Ma by Rb-Sr whole rock isochrons (Bell *et al*., 1977, 1979). The recumbent structures in New World Island are deformed by upright, NE-SW folds that deform dacitic dykes correlated with the Loon Bay Pluton dated at 372 ± 10 and 379 ± 10 Ma by K-Ar (Kay, 1976; Van der Pluijm, 1986), although a latest Silurian age has also been reported for the Loon Bay pluton (Elliott *et al*. in Dunning *et al*., 1990) . In central Newfoundland, the Stony Lake volcanic rocks are inferred to lie unconformably upon cleaved Botwood Group, and thus the 423 +3/-2 Ma U-Pb zircon age from a Stony Lake rhyolitic tuff provides an upper limit on the local deformation in this part of the Exploits subzone (Colman-Sadd and Russell, 1982; Dunning *et al*., 1990). In the northern Mirimichi Highlands, an upper limit on at least one phase of deformation is provided by the oldest K-Ar cooling age on muscovite from a polydeformed granite intrusive into the basal Tetagouche Group (432 ± 17 Ma recalculated from Wanless *et al*., 1973; Helmstaedt, 1973), and the presence of pebbles with a pre-depositional cleavage derived from the Tetagouche Group in Ludlovian conglomerates (Helmstaedt, 1971).

A Middle Silurian, upper limit on the time of deformation in the Notre Dame and Dashwood subzones is provided by the age of unconformably overlying units: 429 +6/-5 Ma U-Pb zircon age of the Springdale Group (Chandler *et al*., 1987); 429 ± 3 and 427 ± 3 Ma U-Pb zircon ages on volcanic units of the Topsails Igneous Suite (Whalen *et al*., 1987); and 429 +7/-3 Ma U-Pb zircon age of the Bear Pond rhyolite (Dunning *et al*., 1990), and post-tectonic plutons: 431 ± 2 and 435 +6/-3 Ma U-Pb zircon ages of the Main Gut gabbro and Boogie Lake monzonite (Dunning *et al*., 1990); and 421 ± 7 Ma Rb-Sr whole rock isochron age of the peralkaline phase of the Topsails granite dykes of which cut across the mylonite zone between the Hungry Mountain Complex and the Buchans Group (Thurlow, 1981; Bell and Blenkinsop, 1981). An upper limit on the high grade structures along the Red Indian Line is also provided by the presence of deformed amphibolite facies pebbles in the middle-late Emsian Windsor Point Group (Kean and Jayasinghe, 1981; Kean, 1983).

Ganderian Episode of Deformation: Silurian - Early Devonian Northwesterly Vergent Deformation Associated with the Acccretion of the Avalon Zone

Kennedy (1976) originally defined the Ganderian Orogeny for the deformation and metamorphic effects that affected the Gander Zone and the neighbouring Dunnage and Avalon Zones. Based on meagre evidence, he inferred that this orogenic event was Late Precambrian in age. Since then, much geochronological data has been published that indicates that this event is mainly Silurian. Thus, the Ganderian episode of deformation is redefined as mainly Silurian deformation found mainly in the Gander Zone and adjacent parts of the Dunnage and Avalon Zones (Figs. 4-10).

A Late Ordovician (Caradocian) lower limit on the time of earliest deformation in the southeastern Exploits subzone is provided by the ages of units involved in the deformation: (1) Davidsville Group: late Arenig - early Caradoc (Bergstrom *et al*., 1974; McKerrow and Cocks, 1977; Stouge, a980; Neuman, 1984); (2) Bay d'Espoir Group: late Arenig - early Caradoc (Neuman, 1984); (3) Bay du Nord Group: 466 ± 3 Ma U-Pb zircon age (Dunning *et al*., 1990), and (4) Caradocian fossils in the upper two thrust slices in the northern Mirimichi Highlands (van Staal, 1988b). In the Gander Zone, the lower limit on the time of deformation is given by the age of the youngest units involved in the deformation: (1) Indian Bay Formation: late Arenig (Wonderley and Neuman, 1984); (2) limestone conglomerate (?)unconformably overlying the Spruce Brook

Formation: late Arenig - early Caradoc (Neuman, 1984; Dec and Colman-Sadd, 1990); (3) quartzite and phyllite in the northern Mirimichi Highlands: late Arenig - early Llanvirn (van Staal, 1988b); and (4) Cookson Formation in southern New Brunswick and adjacent parts of Maine: Tremadocian (-?Arenigian) grapolites (J. Riva, personal communication in Ruitenberg and Ludman, 1978; Cumming, 1967). In the Avalon Zone, a Lower-Middle Ordovician, lower limit for the time of earliest deformation is given by the age of the youngest units affected by the deformation: (1) Wabana Group in Newfoundland: Arenig (Hayes, 1915; Ranger, 1979); (2) mudstones in the Phalarope P62 well on the Grand Banks of Newfoundland: Llanvirn-Arenig (Jenkins, 1984); (3) MacLeod Brook Formation: Arenig (Hutchinson, 1952); and (4) Ferrona Formation: early Ordovician (Williams, 1914; Keppie and Murphy, 1988).

Some Early-Middle Silurian units (La Poile Group, Money Point Group, Ingonish Island volcanics, Sarach Brook rhyolite and Jumping Brook complex) appear to be syn-tectonic deposits that were deformed in the Silurian. The age of these units provides a lower limit on the time of their deformation: (1) La Poile Group: 428 ± 6, 424 +7/-3, and 420 + 8/-2 Ma (U-Pb igneous zircon: Dunning *et al.*, 1990); (2) Money Point Group: 427 ± 4 Ma (U-Pb igneous zircon: Keppie *et al.*, in press); (3) Ingonish Island volcanics: 412 ± 15 Ma (Rb-Sr whole rock isochron: Keppie *et al.*, 1986); (4) Sarach Brook rhyolite: 433 +7/-4 Ma (U-Pb igneous zircon: Dunning *et al.*, 1990); (5) feeder dyke to the Jumping Brook complex: 439 ± 7 Ma (U-Pb igneous zircon age: Currie *et al.*, 1982).

Various stages of Silurian time of deformation are provided by ages of syn- to post-tectonic plutonism and cooling ages following dynamothermal metamorphism. Thus, syn- to post-D$_1$ stages range from ca. 430 to 420 Ma based on: (1) the anatectic Lockers Bay granite, which cuts the gneissic foliation in the Gander Zone (Hare Bay gneiss) and contains migmatitic gneiss xenoliths polydeformed prior to incorporation, yielded a 460 ± 20 Ma U-Pb zircon intrusive age (Dallmeyer *et al.*, 1981); (2) syntectonic Cape Freels pluton: 414 ± 5 Ma Rb-Sr whole rock isochron (Bell *et al.*, 1977); (3) syntectonic foliated granodiorite along the western side of the Ackley pluton has yielded a Rb-Sr whole rock isochron age of 427 ± 15 Ma (Tuach and Kontak, 1986); (4) the Little Passage Gneiss yielded a 423 +5/-2 Ma U-Pb metamorphic zircon age (Dunning *et al.*, 1990); (5) the syn-D$_1$, Gaultois Granite yielded a 421 ± 2 Ma U-Pb igneous zircon age (Dunning *et al.*, 1988); (6) low grade phyllite from the Through Hill area yielded a slightly discordant ^{40}Ar/^{39}Ar age spectrum with an intermediate plateau age of 409 ± 6 Ma (Dallmeyer *et al.*, 1983), interpreted as the age of the cleavage; (7) various phases of the Burgeo Intrusive Suite that stitches the boundary between the Exploits and Avalon bracket the early phase of deformation: syn-D$_1$ granodioritic phase of the Burgeo Intrusive Suite yielded a 429 +5/-3 Ma U-Pb zircon age and the post-D$_1$, pre-D$_2$ granitic phase of the Burgeo Intrusive Suite yielded 415 ± 2 Ma (Dunning *et al.*, 1990); (8) the post-D$_1$ La Poile pluton gave a concordant igneous zircon age of 416 ± 4 Ma (Chorlton and Dallmeyer, 1986); (9) the syn-D$_2$ phase of the North Bay granite suite yielded 430 ± 4 and 427 ± 12 Ma Rb-Sr whole rock isochrons ages (Elias and Strong, 1982; O'Brien *et al.*, 1988); (10) a syn- to post-D$_1$ granite in central Mirimichi Highlands yielded a 432 ± 6 Ma Rb-Sr whole rock isochron age (Fyffe *et al.*, 1977).

Cooling through ca. 650-550°C temperatures ranged between ca. 420 and 410 Ma in Newfoundland: it is recorded by U-Pb titanite ages: (1) the syn-tectonic Gaultois granite: 417 ± 2 Ma U-Pb titanite age (Dunning *et al.*, 1990); (2) 417 ± 2 Ma U-Pb titanite age from the earliest syn-D$_1$ phase of the Burgeo Intrusive Suite (Dunning *et al.*, 1990); (3) the upper amphibolite facies Port aux Basque Complex yielded 412 ± 2 Ma metamorphic titanite ages (Dunning *et al.*, 1990).

Cooling through ca. 500°C took place between ca. 400 and 380 Ma and is recorded by ^{40}Ar/^{39}Ar hornblende cooling ages: (1) the 397 ± 5 Ma age in the Hare Bay gneiss (Dallmeyer *et al.*, 1981); (2) post-D$_2$ ^{40}Ar/^{39}Ar hornblende plateau cooling ages of 378 ± 5 to 388 ± 5 Ma in the Bay du Nord Group (Chorlton and Dallmeyer, 1986); (3) 385-380 Ma ^{40}Ar/^{39}Ar hornblende ages in amphibolite facies rocks of the Money Point Group in northern Cape Breton Island (Keppie *et al.*, in press); (4) ca. 384-423 Ma ^{40}Ar/^{39}Ar hornblende total gas ages in the central Cape Breton Highlands (Reynolds *et al.*, 1989).

Cooling through ca. 350-300°C took place between ca. 385 and 360 Ma and is recorded by Rb-Sr and ^{40}Ar/^{39}Ar muscovite and biotite ages: (1) the 383 ± 6 to 365 ± 6 Ma biotite ages in the Hare Bay gneiss (Dallmeyer *et al.*, 1981); (2) biotite total-fusion ^{40}Ar/^{39}Ar cooling ages of 410 ± 4 Ma and 392 ± 8 Ma in the western part of the Ackley pluton (Kontak *et al.*, 1988); (3) ^{40}Ar/^{39}Ar plateau ages on biotite of 395 ± 5 Ma and 389 ± 5 Ma in Newfoundland (Dallmeyer *et al.*, 1983); (4) ca. 365-380 Ma ^{40}Ar/^{39}Ar muscovite plateau ages from metamorphic and plutonic rocks of the northern Cape Breton Island (Keppie *et al.*, in press); (5) ca. 430-360 Ma ^{40}Ar/^{39}Ar biotite total gas ages (Reynolds *et al.*, 1989); (6) a post-D_1 foliated granite intruding the Tetagouche Group yielded a K-Ar muscovite age of 432 ± 17 Ma (recalculated from Wanless *et al.*, 1973).

Late syn-tectonic to post-tectonic plutons range in age from ca. 420 to 390 Ma with cooling ages as young as ca. 370 Ma: (1) post-tectonic Middle Brook granite: 435 ± 22 Ma Rb-Sr whole rock isochron (recalculated from Bell and Blenkinsop, 1875; Blackwood, 1977); (2) the main part of the Ackley pluton, dated between 378 and 374 ± 5 Ma (total-fusion ^{40}Ar/^{39}Ar ages on hornblende, muscovite and biotite; Kontak *et al.*, 1988), stitches the Gander-Avalon Zone boundary and so provides an upper limit on all the deformation; (2) the post-tectonic Through Hill granite yielded a 429 ± 2 Ma Rb-Sr whole rock isochron (Elias and Strong, 1982; Colman-Sadd and Swinden, 1984); (3) the youngest post-tectonic phase of the North Bay granite suite yielded a 396 +6/-3 Ma U-Pb igneous zircon age (Dunning *et al.*, 1990); (4) the post-D_1 Overflow Pond granite yielded a U-Pb upper intercept age of 387 ± 8 Ma (Dallmeyer *et al.*, 1983); (5) the post-tectonic Chetwynd granite yielded a 390 ± 3 Ma igneous zircon age (Dunning *et al.*, 1990), and ca. 390-370 Ma ^{40}Ar/^{39}Ar hornblende and biotite plateau cooling ages (Chorlton and Dallmeyer, 1986); (6) foliated Neils Harbour orthogneiss on the northeast coast of the Cape Breton Highlands has yielded a 403 ± 3 Ma concordant igneous zircon age (Dunning *et al.*, 1990); (7) post-tectonic Cameron Brook granodiorite in the eastern Cape Breton Highlands has yielded a 402 ± 3 Ma U-Pb zircon age (Dunning *et al.*, 1990); (8) foliated Cheticamp Lake orthogneiss in central Cape Breton Highlands yielded a 396 ± 2 Ma U-Pb concordant igneous zircon age (Dunning *et al.*, 1990); (9) post-tectonic Mount Elizabeth pluton in the northern Mirimichi Highlands yielded a 417 ± 2 and 414 +11/-1 Ma U-Pb monazite and zircon ages, respectively (Bevier and Whalen, 1990); (10) late tectonic North Pole granite yielded a 417 ± 1 Ma U-Pb monazite age (Bevier and Whalen, 1990); (11) various phases of the late- to post-tectonic Pokiok batholith yielded U-Pb ages ranging from 415 ± 1 to 402 ± 1 Ma U-Pb titanite and monazite ages (Bevier and Whalen, 1990); (12) the late tectonic Junipers Barren granite yielded Rb-Sr and K-Ar muscovite ages of 433 ± 4 Ma to 421 ± 6 Ma, respectively (Bevier and Whalen, 1990); (13) various other granitic plutons in the Mirimichi Highlands have yielded Rb-Sr and K-Ar muscovite ages ranging from 422 ± 4 to 408 ± 7 Ma (Bevier and Whalen, 1990); (14) the post-tectonic Tower Hill pluton is intrusive into the Fredericton Trough was dated as 412 ± 15 Ma by a Rb-Sr whole rock isochron (Butt, 1976; Fyffe *et al.*, 1981; Bevier, 1988); (15) the St. George Batholith cuts structures in the Fredericton Trough and has yielded U-Pb zircon ages ranging from 430 ± 3 to 366 ± 1 Ma (Bevier, 1989), and the western part of the batholith has yielded and a 406 ± 7 Ma Rb-Sr whole rock isochron Fyffe *et al.*, 1981); (16) post-tectonic St. Stephen pluton that contains cleaved sedimentary xenoliths and is intruded into the Cookson Formation, yielded 420 to 406 ± 20 and 370 ± 16 Ma K-Ar hornblende and biotite ages, respectively (Wanless *et al.*, 1973); (17) late tectonic Pabineau pluton dated at 394 ± 1 Ma (U-Pb zircon age, Bevier, 1988; van Staal, 1987). Thus, most of the deformation is Silurian with some low temperature deformation lasting into the Early Devonian.

Stratigraphic constraints provide an upper limit on structures in specific areas: (1) in the northern Antigonish Highlands, Llandoverian Beechhill Cove Formation rests unconformably upon polyphase structures in the Cambro-Ordovician rocks (Keppie and Murphy, 1988); (2) in southern New Brunswick the late Llandoverian C3 to Wenlockian Waweig Formation unconformably overlies polydeformed Cookson Formation (Boucot *et al.*, 1966; Berry and Boucot, 1970).

INTERPRETATION

The earliest deformation recorded in the southeastern part of the Central Mobile Belt is Early Ordovician (Tremadocian or Arenigian: assigned to the Penobscotian) and occurs in the GRUB ophiolites and associated ophiolitic melange. It may be related to the obduction of oceanic lithosphere of the Exploits subzone onto the Gander Zone. Unfortunately, kinematic data are not presently available, so the direction of obduction is uncertain. The short time interval (<15 Ma) between the formation of the oceanic lithosphere and its deformation suggests that it was relatively hot and buoyant when it was obducted.

This Early Ordovician deformation is presumably related to subduction. Subduction-related volcanic arc rocks varying from island arc tholeiites through calcalkaline basalts to back-arc units range in age from 513 ± 2 through 498 +6/-4 to 462 +4/-2 Ma (U-Pb igneous zircon ages: Evans *et al.*, 1990) occur in the Victoria Lake Group in central Newfoundland. This suggests that some subduction preceded the docking and amalgamation of the Exploits subzone and the Gander Zone. The calc-alkaline volcanics of the Sleepy Cove Group on Twillingate Island may represent a continuation of this volcanic arc (Fig. 18). These data suggest a NW-polarity for subduction. The early Ordovician deformation recorded in Twillingate could be due to transcurrent or transform movements within the volcanic arc. On the other hand, the early Ordovician deformation in the Miramichi Highlands could be attributed to collision of this region with the trench (Fig. 18).

Given that most of Newfoundland is now underlain by continental lithosphere (Karlstrom, 1982), and assuming that the ophiolites were rooted in the "suture" identified on the reflection seismic profile (Keen *et al.*, 1986) the minimum amount of transport in a SSE direction for the Pipestone Pond and GRUB ophiolites is ca. 100 km. Note that this does not take into account any transport of the suture itself. To this may be added a further ca. 100 km if the sole thrust beneath the ophiolites is correlated with the Day Cove thrust (Colman-Sadd and Swinden, 1984). The extent of telescoping on other thrusts in the southeastern part of the Central Mobile Belt is presently unknown, however it could be conservatively estimated to be similar to the ca. 200 km on the sub-ophiolite thrust giving a total of >400 km. This value is closely comparable to that derived above for the Taconian on the northwestern side of the orogen. Assuming that most of this movement on the thrust occurred in the Early Ordovician (Tremadoc-Arenig: ca. 25 Ma), then the rate of convergence across the southeastern part of the Central Mobile Belt is >1.6 cm/year.

The Late Ordovician - Early Devonian deformation centred along the axis of the Dunnage Zone (Beothukan) appears to be related to SE- to SSE-thrusting and nappe emplacment, although locally NW-thrusting has also been recorded. This may be related to amalgamation of the Notre Dame and Exploits subzones from which it follows that subduction took place between them. The subduction-related, magmatic arc, Middle to Late Ordovician igneous rocks in the Dashwood and eastern Notre Dame subzones (Dunning and Chorlton, 1985; van Berkel *et al.*, 1986) indicates that the Notre Dame/Dashwood subzones lay on the upper plate, and that ca. 20 Ma of subduction took place before the arrival of the Exploits subzone at the trench in the Caradocian. Accretionary deformation between these two subzones continued throughout the Silurian and into the Devonian. A possible extension of associated subduction may occur in southern Eire where the late Ordovician-early Devonian volcanic rocks are calc-alkaline (Stillman and Francis, 1979).

The Silurian - Early Devonian deformation (Ganderian) centred on the Gander Zone and adjacent Dunnage and Avalon Zones may be related to the sinistral accretion of the Avalon Zone approximately parallel to the NE-trend of the orogen. The variation from transcurrent shear deformation along NE-trending structures to thrusting along NW-trending zones may be related to irregularites in the western boundary of the Avalon Zone. Structures in the Burgeo Terrane, a promontary of the Avalon Zone, indicate that here the Avalon Zone was obducted over the Central Mobile Belt. Subduction-related volcanic arc rocks of Early-Middle Silurian age occur in the Avalon Zone of northern Cape Breton Island (Macdonald and Smith, 1980; Jamieson *et al.*, 1990; Keppie *et al.*, in press). The late Ordovician deformation associated with steeply dipping shear zones in the northern Antigonish Highlands and southern New Brunswick may be early manifestations of this accretionary event. This is consistent with the paleomagnetic data which suggests that a sinistral megashear existed between the Central Mobile Belt and the Avalon Zone in the Silurian and Early Devonian (Kent and Keppie, 1989). These paleomagnetic data suggest that at least 1500 km relative sinistral displacement took place during Silurian and early Devonian times (ca.40-50 Ma) and gives a relative rate of motion of ca. 3.75-3.0 cm/year. Kent and van der Voo (1990) have suggested that this coincided with collision between Laurentia and Gondwana (South America).

Acadian and Alleghanian Episode of Deformation: Early Devonian - Mid Permian Affecting All Zones

The term, Acadian Orogeny, is derived from the name given to French immigrants in Maritime Canada. Its geological usage has traditionally been for the Early-Middle Devonian Orogeny recorded in the rocks of this same area. However, over the years the time span of the term has been expanded from the Silurian through to the end of the Devonian and structures have been assigned to it from right across the Appalachian Orogen in Canada. Late Devonian deformation in the central and southern Appalachians has been genetically linked with the Late Devonian clastic wedge (Quinlan and Beaumont, 1984). Its usage herein is restricted to Devonian deformation affecting the entire orogen. The terminal stages of the Beothukan and Ganderian deformational episodes probably overlap the onset of the Acadian deformation making distinctions difficult. A distinction may be made on genetic criteria as deformation changed from predominantly sinistral during accretion of the Avalon Zone to mainly dextral associated with accretion of the Meguma Zone.

The term, Alleghanian Orogeny, was used by Woodward (1957) for the deformation which produced the NW-vergent folds and thrusts in the Paleozoic rocks (up to and including the Dunkard Group: Late Pennsylvanian or Early Permian) of the Valley and Ridge and adjacent Cumberland Plateau in the U.S. Appalachians. This was genetically related to the deposition of Carboniferous and (?)Permian clastic wedges in the foreland basin, the bases of which are marked by the upward transition from platformal limestone to shale in the Late Mississippian (Ferm, 1974; Quinlan and Beaumont, 1984). This has recently been linked to synchronous deformational events in the overthrust crystalline allochthons (Secor *et al.*, 1986a and b; Dallmeyer *et al.*, 1986). Recent geochronological work has shown that W-vergent accretion of the Meguma Terrane started in the Early Devonian and continued periodically until middle Permian times (Keppie and Dallmeyer, 1987; Dallmeyer and Keppie, 1987). This suggests that the Acadian and Alleghanian deformation may be genetically linked.

The onset of Acadian deformation is diachronous from Early Devonian in the southeast to latest Devonian in the northwest. Accordingly, the following description of the Acadian structures will be from southeast (Meguma Terrane: uppermost allochthon) to northwest on the foreland. Except for the Meguma Terrane, for which there is much age data, emphasis is placed upon structures present in the Late Paleozoic rocks where the age of the structures is unequivocal. Their presence in Lower Paleozoic and Precambrian rocks may usually only be inferred by comparison with structures in the Upper Paleozoic rocks.

The Meguma Terrane is bounded on its northern side by a major E-W shear/fault zone (the Minas Fault/Geofracture) with dextral motions of late Paleozoic age followed by sinistral movements in the Mesozoic (Keppie, 1982). Traced westwards, this boundary swings into the NE-SW Bay of Fundy (Fig. 1). Seismic profiles in the Bay of Fundy show the major structure to be a SE-dipping Mesozoic listric normal fault inferred to be located along a late Paleozoic listric thrust zone (Brown, 1986; Keen *et al.*, 1991). Seismic data allows two interpretations for the down-dip extension of the sole thrust: either it maintains its SE dip to the Moho or it flattens into a zone between the lower and upper crusts at a depth of about 15 km (Keen *et al.*, 1991). The latter alternative is supported by the recovery of lower crustal xenoliths with Ordovician model Sm/Nd ages in late Devonian mafic dykes cutting the Meguma Group along the southern coast of Nova Scotia (Eberz *et al.*, in press). This implies that parts of the lower crust are younger than the overlying Cambrian-Ordovician Meguma Group, and may only be explained by overthrusting the Meguma Zone. Thus, the Minas Fault may be interpreted as a lateral ramp to the sole thrust beneath the Meguma Zone.

The Late Paleozoic rocks of the Maritimes Basin have been subdivided into three upward-fining, generally unconformity-bounded megasequences that appear to be common to most of the region: Late Devonian to Early Namurian, Late Namurian to Westphalian A and Westphalian B to Early Permian (Ryan *et al.*, 1987; Gibling *et al.*, 1987). A fourth megasequence may be represented by fragmentary Early to Middle Devonian rocks. These megasequences may be related to tectonic activity that was generally more active during early stages of each megasequence. The Middle Devonian, Late Devonian to Early Namurian and

Late Namurian - Westphalian A megasequences all start with local, rift-related volcanism (Dostal *et al.*, 1983; Wilton, 1983; Payette and Martin, 1986; Fyffe and Barr, 1986). A proximal source for the detritus is inferred for the Devonian and Early Carboniferous megasequences, whereas a distal source is postulated for the uppermost megasequence: the Late Namurian - Westphalian A megasequence has both local and more distal sources (Ryan *et al.*, 1987; Gibling *et al.*, 1987). This change in the source region from proximal to distal is accompanied by an increase in basin size from small in the Devonian and Early Carboniferous to extensive in the Late Carboniferous and Permian. The structure in the Late Paleozoic rocks is dominated by a series of NE-trending faults and associated folds which merge with the W-trending Minas Fault that bisects Nova Scotia. At its western end the Minas Fault swings southwest and its late Paleozoic character changes from a dextral strike-slip fault to an oblique thrust (Keppie and Dallmeyer, 1987). Several other E-W faults show similar swings in trend and the changes in the nature of the faulting, e.g. Grand Pabos-Restigouche, McKenzie Gulch, Rocky Brook-Millstream, Catamaran, Mira-Bateson, Gunflap-Cape Ray faults (Anderson, 1972). Normal, reverse and strike-slip motions have been documented on the NE-trending faults. A few NW-trending faults are also present, e.g. the Round Valley Fault in the southern part of the Bay St. George basin (Knight, 1983) separates complexly folded rocks to the southwest from a monoclinally dipping succession to the northeast explicable in terms of differential shortening on either side of the fault during the folding. In general, structural complexity generally decreases with decreasing age.

AGE OF STRUCTURES

The polygenetic history of many of the faults in the Late Paleozoic rocks makes it difficult to determine a coherent picture of their evolution through time. However, there are some specific data to indicate that the structures may be described in association with the stratigraphic megasequences.

Early-Middle Devonian deformation

The Lower Paleozoic rocks of the Meguma Zone are deformed by major upright-asymmetric folds which swing from N-S in the southwest through NE-SW to E-W in the east adjacent to the Minas Fault (Fig. 1) (Keppie, 1976, 1979). These folds are associated with a spaced cleavage in the psammitic rocks which refracts into a slaty cleavage in the pelitic lithologies. The en echelon nature of some folds and the nonparallelism of minor structures in some places led Keppie (1987) to infer that they were produced in a sinistral shear regime, while in other locations parallelism of the minor structures suggested orthogonal shortening (Henderson *et al.*, 1986). A lower limit on the age of these structures is the late Seigenian (-?early Emsian) age of the Torbrook Formation (Boucot, 1960), the youngest deformed unit. An upper limit is provided by (i) the Tournaisian age of the lowest unconformably overlying unit of the Horton Group along the northern margin of the Meguma Zone (Howie and Barss, 1974); and (ii) the 374 ± 2 Ma U-Pb (Keppie *et al.*, 1985) and 372 ± 2 Ma Rb-Sr whole rock isochron (Clarke and Halliday, 1980) ages of the oldest crosscutting granitoid plutons. The age of the slaty cleavage has been dated using the ^{40}Ar/^{39}Ar incremental release technique on whole rock phyllites and gave ages of 405 ± 7 Ma, ca. 395-400 Ma, ca. 385-389 Ma, and 377-382 Ma in discrete strike parallel belts (Reynolds and Muecke, 1978; Keppie and Dallmeyer, 1987; Reynolds *et al.*, 1987) indicating diachronous development over ca. 30 Ma.

The Middle-Late Devonian rocks generally rest unconformably upon a variety of previously deformed, older rocks. Deformation associated with the Middle Devonian megasequence is best documented in the Antigonish Highlands where NE-SW folds with a subvertical slaty cleavage in the Arisaig Group (including the Siegenian Knoydart Formation) are unconformably overlain by gently dipping unfossiliferous McAras Brook Formation that is conformably overlain by rocks containing late Eifelian to early Givetian spores (Boucot *et al.*, 1974; Keppie *et al.*, 1978; D. C. MacGregor, written comm.). The en echelon arrangement of folds in the Arisaig Group and their anticlockwise orientation relative to the NE-trending Hollow Fault imply dextral motions on the Hollow Fault at this time (Boucot *et al.*, 1974; Yeo and Ruixiang, 1987). In the Fredericton Trough, Silurian-Lower Devonian (upto Gedinnian Eastport = Flume Ridge Formation) were deformed prior to intrusion of the St. George Batholith, the western part of which has yielded a 395 ± 2 Ma U-Pb zircon age (Butt, 1876; Fyffe *et al.*, 1981; Bevier, 1988). In the southern Miramichi Highlands, Siluro-Devonian rocks (as young as the late Emsian Wapske Formation) were folded and cleaved before intrusion of the Skiff

Lake granite (Lutes, 1979; Crouse, 1977, 1979; 1981 a and b; St. Peter 1980, 1982) dated at 409 ± 2 Ma (U-Pb zircon age; Bevier, 1988) and 389 ± 20 Ma (Rb-Sr whole rock isochron; Fyffe *et al.*, 1981).

In the northern New Brunswick and Gaspe Peninsula, evidence for deformation in Late Silurian to Middle Devonian times is limited to local unconformities/disconformities (Fig. 5). Thus, a local unconformity is present beneath the Indian Point Formation of Gedinnian - (?)Pridolian age (Bourque and Lachambre, 1980). The hiatus varies from zero to most of the Ludlovian and Pridolian. This break in Maine was termed the Salinic Disturbance by Boucot (1962). Local unconformities are also present between (i) the Emsian La Garde Formation and the early Siegenian - late Silurian Restigouche Formation; (ii) the Emsian/Siegenian Grande Greve and Siegenian Cap Bon Ami Formation; (iii) the Emsian York River and the Emsian/Siegenian Grande Greve Formations; and (iv) the Fleurant Formation (overlain conformably by the Frasnian Escuminac Formation) and the Emsian Pirate Cove Formation (St. Julien *et al.*, 1972; Brideaux and Ridforth, 1970; Boucot, 1968; Dineley and Williams, 1968 a and b; Rust, 1981, 1982). These unconformities appear to be about the same age as the unconformities recognized in Maine which have been correlated with a phase of folding (Boucot *et al.*, 1964). However, it should be noted that in Maine the overlying units (middle Devonian Trout Valley and Mapleton Formations) are also folded and it is thus uncertain how intense was the early Devonian deformation. Certainly, the main deformation of equivalent rocks in Canada is post-Escuminac Formation (Frasnian) and so may be related to the Acadian.

Late Devonian - Early Namurian deformation

In the Meguma Zone, there are several discrete dextral shear zones or fault zones (Fig. 11). Dynamically recrystallized mica from these shear zones gave $^{40}Ar/^{39}Ar$ plateau ages which are different for each shear zone: (i) 359-372 Ma; (ii) 350-356 Ma; (iii) 315-325 Ma; (iv) 285-295 Ma; and (v) 270 ± 8 Ma (Keppie and Dallmeyer, 1987; Dallmeyer and Keppie, 1987). The close similarity between the muscovite and biotite plateau ages together with the proximity of the sub-Carboniferous unconformity to the present erosion surface indicate rapid cooling through ca. 300° C and suggests these dates closely post-date the ages of active shearing. The magnitude of displacements along the intra-Meguma shear zones is probably limited because, even though some shear zones cut the South Mountain Batholith its boundaries are not significantly offset. It is concluded that the Meguma Zone was in a dextral shear regime throughout the Carboniferous and early Permian, the location of the active shear zone shifting with time, although most movement appears to be concentrated along the Minas Fault.

Synchronous deformation of the Carboniferous rocks (Fig. 1) produced: (i) en echelon ENE-WSW folds in the Tournaisian Horton Group along the northern side of the Meguma Terrane (Keppie, 1979); (ii) recumbent folds and thrusts refolded by upright ENE-WSW folds in the Viséan Windsor Group (Fyson, 1967; Giles and Boehner, 1982; Moore, 1967; Moore and Ferguson, 1986); and (iii) E-W to ENE-WSW fault zones. The Westphalian C/D Scotch Village Formation appears to post-date the recumbent folding, however it is affected by ENE-WSW, upright folds and is locally overturned adjacent to some faults (Stevenson, 1959; Keppie, 1979). Kinematic analysis of the ENE-WSW upright, en echelon folds and associated faults indicates a dextral sense of motion on the E-W St. Mary's Graben.

The Late Devonian to Early Namurian megasequence (0->3 km thick) was deposited in a series of small basins bounded by NE-trending faults that vary from half graben (generally N-thickening sedimentary wedges: e.g. Sydney and Minas basins; occassionally S-thickening sedimentary wedges: e.g. Moncton, Bay St. George and Deer Lake basins) to graben (e.g. Cumberland Basin, Sackville Sub-basin)(Carter and Pickerill, 1985; Gibling *et al.*, 1987; Martel, 1987). Some of these basins are bounded by faults with other trends, e.g. NW-trending Port Elgin Fault, N-trending Memramcook River Fault (Martel, 1987). These authors have documented facies changes from alluvial fans adjacent to NE-trending faults to fluvial, lacustrine and marine sediments towards basin centres for the Late Devonian-Tournaisian parts of many of these small basins, indicating active faulting during deposition. East and north of the Caledonian Highlands, Gussow (1953) documented the presence of both NW- and SE-vergent thrusts and associated NE-trending folds. Some of these structures are truncated by the unconformity at the base of the Enrage Formation (?late Viséan - late Namurian) (Gussow, 1953;

McLeod, 1979) indicating that some of this deformation is pre-Late Carboniferous. Particularly significant are the faults bounding the northwestern margin of the Caledonian Highlands, which all appear to be NW-vergent thrusts. Carter and Pickerill (1985) have shown that the Late Devonian - Tournaisian Horton Group was deposited in an asymmetric basin thickening and coarsening southwards towards the northern margin of the Caledonian Highlands suggesting active NW-thrusting throughout this time interval. The upward coarsening and northward migration of the coarser Horton facies is consistent with NW-thrusting of the Caledonian Highlands during the Tournaisian.

In Newfoundland, alluvial fans on the northwestern side of the Long Range/Grand Lake Faults indicates active faulting during deposition of the (?)Late Devonian Kennels Brook Formation, the Tournaisian Anguille Group, and the Visean Codroy Group (Belt, 1969; Hyde, 1979; Knight, 1983). Belt (1969) and Knight (1983) also inferred that an E-W normal fault with a southerly downthrow was active during deposition of the Anguille Group in order to explain the absence of many of the units north of the fault. Late Paleozoic rocks are rarely preserved elsewhere in Newfoundland: along the Cape Ray Fault as the Emsian-Eifelian Windsor Point Group (Chorlton, 1978, 1980, 1983); around Red Indian Lake as the Shanadithit Formation (Kean, 1977); and in the Avalon Zone as a number of rock units (Greene, 1975). Along the NE-SW portion of the Cape Ray Fault Zone, Chorlton (1983) and Wilton (1981) deduced that the mylonites and shear zones of post-Windsor Point age involved reverse faulting of east over west. Where the Cape Ray Fault Zone swings into the E-W Gunflap Hills Fault, the movement was transformed into oblique thrusting with a dextral shear component. This uplift is reflected in the ^{40}Ar/^{39}Ar plateau ages in hornblende of 361-380 ± 5 Ma and in biotite of 343-360 ± 5 Ma (Chorlton and Dallmeyer, 1987) in the area southeast of the Cape Ray/Gunflap Hills Faults. The last structures to affect the Windsor Point Group are chevron folds, kink bands and crenulation cleavage. Farther north around Red Indian Lake, the Shanadithit Formation is either undeformed or only gently folded and faulted (Kean, 1977).

In western Newfoundland, the Taconian Humber Arm allochthon also displays the effects of Devonian deformation (Cawood and Williams, 1988). While Taconian deformation is W-vergent and confined to the allochthon, Devonian deformation extends into the miogeoclinal rocks and the Grenvillian basement in the form of both W- and E-vergent thrusts and folds. Westward basement wedging is inferred to have caused delamination of cover from basement resulting in eastward backthrusting. The age of some of this deformation is constrained between the Early Devonian age of the youngest deformed rocks and the undeformed Carboniferous rocks (Cawood and Williams, 1988).

In the Avalon Zone of Newfoundland, the Upper Devonian Great Bay de l'Eau Formation has been involved in SE-vergent thrusting (Greene, 1975). Synchronous movements along the NE-trending Paradise Sound Fault may be indicated by the ^{40}Ar/^{39}Ar plateau ages on phyllites ranging from 380 to 353 ± 10 Ma (Dallmeyer *et al.*, 1983). While most of the deformation along the Dover-Hermitage Fault appears to be pre-late Devonian, continued deformation may be indicated by the ^{40}Ar/^{39}Ar plateau ages on biotite ranging between 383 and 365 ± 6 Ma and the discordant phyllite ages of 309 to 352 ± 10 Ma from samples collected adjacent to the fault (Dallmayer *et al.*, 1981).

Waning fault movements during deposition of the Visean Windsor Group is suggested by its overstep across faults, although some continued activity is indicated by local facies changes (e.g. Bradley and Bradley, 1986; Gibling *et al.*, 1987). Sediments are generally fine grained in the Early Namurian Canso Group. The earliest motion along the Quaco Head Fault in southern New Brunswick is indicated by the coarsening of the basal Hopewell Group conglomerates of late Visean or early Namurian age adjacent to the fault (Plint and Van de Poll, 1984).

In the Gaspe Peninsula, the major structure consists of a train of W-trending folds which swing into the SW-regional trend when traced inland (Keppie *et al.*, 1982). A series of N-vergent thrusts run parallel to these folds, and have been traced offshore passing south of Anticosti Island (Roksandic and Granger, 1981). In places these thrusts may be observed swinging into NW-SE dextral faults and NE-SW sinistral faults (McGerrigle, 1950; Brisebois, 1981; Beland, 1978), which may be interpreted as conjugate tear faults associated with the N-S shortening. These structures deform rocks up to and including the Frasnian Escuminac Formation, but are generally unconformably overlain by the Bonaventure - Cannes de Roche Formation containing Namurian spores (Howie and Barss, 1974). These data bracket the main phase of deformation as post-Frasnian/ pre-Namurian.

Late Namurian - Westphalian A deformation

The Late Namurian - Westphalian A megasequence (0-2.3 km thick) is mainly preserved in northern Nova Scotia and southern New Brunswick, and is absent in the Meguma Zone and in the Sydney Basin (Boehner and Prime, 1985; Ryan *et al.*, 1987). Some of the NE-trending faults limit the deposition of the lower parts of this megasequence, however, the upper parts generally overstep the faults. Most anticlines and associated salt diapirs in the Cumberland Basin and southern New Brunswick are unconformably overlain by the Westphalian C basal Pictou Group, which places an upper limit on their development and associated fault movements (Ryan *et al.*, 1987). Traced westwards, the W/WSW-trending folds and faults in the Cumberland Basin in Nova Scotia swing into SW-trending structures in southern New Brunswick (Keppie *et al.*, 1982). This transition is best observed across Shepody Bay, where E-W, upright-asymmetric folds in the Maringouin Peninsula become NW-vergent folds associated with SE-dipping thrusts (e.g. Harvey-Hopewell and Locher Lake Faults) between Shepody and Salisbury Bays (Gussow, 1953). These structures affect rocks as young as the early Westphalian Boss Point Formation, but are unconformably overlain by the late Westphalian Pictou Group.

Alluvial fans adjacent to the Long Range/Grand Lake Faults, indicative of active fault movements, have been recorded in the Namurian - Westphalian C Barachois and Deer Lake Groups (Belt, 1969; Hyde, 1979; Knight, 1983).

In the Gaspé Peninsula, the Namurian Cannes de Roche and Bonaventure Formations are interpreted as alluvial fan deposits (Rust, 1981) indicative of deposition close to active faults. That some of the deformation continued after the Namurian is shown by the fact that the Grand Pabos Fault cuts the Bonaventure Formation (Ayrton, 1967).

Westphalian B - early Permian deformation

The Westphalian B - Early Permian megasequence begins with deposition of the Westphalian B Cumberland Group that is restricted to southern New Brunswick and the western Cumberland Basin. Alluvial fans adjacent to the Cobequid Highlands change to fluvial deposits in the basin centre and indicate active movements along W-trending faults bordering the Cobequid Highlands (Ryan *et al.*, 1987). Similarly, Westphalian A-C Tynemouth Creek and Westphalian (B?) Balls Lake formations along the southern coast of New Brunswick have been interpreted as alluvial fans deposited adjacent to a NE-trending thrust welt to the southeast (Caudill and Nance, 1986; Nance and Warner, 1986). These units were then deformed by NW-vergent folds and thrusts associated with a lower greenschist facies mylonitic cleavage of muscovite and chlorite, overprinted by conjugate NW- and SE-vergent thrusts and folds with a chloritic cleavage and then by NW-trending kink bands (Nance and Warner, 1986). Deposition of the Cumberland Group was followed by Westphalian C - Early Permian Pictou Group which oversteps most of the faults (e.g. Big Pond and Coxheath faults) and has a distal source. A local graben (Stellarton Graben) occurs between the Hollow and Cobequid faults exhibits local alluvial fans adjacent to the faults that grade into deltaic and lacustrine deposits in the basin centre. Kinematic analysis of the faults indicates dextral motions and suggest that the Stellarton Graben originated as a syntectonic rhomb graben during Westphalian C-D time (Yeo and Ruixang, 1987).

Latest Carboniferous to early Permian deformation is documented by structures that deform Westphalian-Permian rocks:

(1) Westphalian C/D Scotch Village Formation is deformed by ENE-WSW, upright folds and is locally overturned adjacent to some faults (Stevenson, 1959; Keppie, 1979).

(2) in the Sydney Basin, NW/N-trending, W-vergent thrusts and associated E/NE-trending folds deform the Westphalian C to Stephanian Morien Group (Boehner and Giles, 1986);

(3) in southern Cape Breton Island, Weeks (1954) has interpreted the Lennox Passage Fault as a NW-vergent thrust associated with recumbent folds, which places lower Carboniferous rocks upon Namurian-Westphalian A rocks: this thrust motion was superimposed upon an earlier listric normal fault (Boehner, 1984);

(4) in western Cape Breton Island, rocks as

young as late Westphalian Broad Cove Formation is deformed by NE-, N-, and NW-trending reverse/thrust faults, most of which have a W-vergence although some show a conjugate sense of vergence (Norman, 1935; Cameron, 1948; Currie, 1977);

(5) in the Antigonish Basin, an allochthon (internally deformed by variably oriented folds and faults) with rocks ranging in age from Viséan Windsor Group to early Westphalian Port Hood Formation ride above a thrust located in evaporites (Boehner and Giles, 1982);

(6) the W-trending Cobequid Fault deforms rocks as young as the Pictou Group and is associated with minor structures and E/ENE-trending folds that indicate a dextral shear sense (Eisbacher, 1969; Donohoe and Wallace, 1985; White, 1983);

(7) in the Cumberland Basin, the Pictou Group is gently folded by E/ENE-trending folds (Ryan *et al.*, 1990);

(8) in southern New Brunswick, Pictou Group is gently deformed by NE-trending folds;

(9) NE-trending faults bounding the Kingston Uplift cut rocks as young as Westphalian C and dip away from the uplift (Gussow, 1953): they appear to have had no influence on the sedimentation (Carter and Pickerill, 1985);

(10) the Catamaran Fault, which cuts Pictou Group in Mirimichi Bay (Howells and McKay, 1977), has a 6 km dextral displacement across the Mirimichi Highlands (Anderson, 1972), but traced westwards it swings into a S/SSE-trending, W-vergent thrust dipping 30-45° east with several SW-trending thrust branches with both NW- and SE-vergence (St. Peter, 1980, 1982): one of these subsidiary faults is unconformably overlain by the Carlisle Formation (conformably overlain by Late Carboniferous Mountain View Formation), while others (Howard Brook, Armond Brook and Woodstock Faults) cut the Carlisle Formation;

(11) in the Plaster Rock outlier, St. Peter (1979) has mapped a series of NW-vergent, NE-trending thrust/reverse faults cutting generally gently dipping rocks of the Arthurette Redbeds (correlated with the Tournaisian Horton Group) closely related to several E-W, vertical dextral faults;

(12) in Newfoundland, thrusts with trends varying from E-W to NE-SW, are mainly SE-dipping and NW-vergent (e.g. east of Deer Lake, the Pynns Brook ophiolitic complex is thrust northwestwards over the Deer Lake Group; Williams *et al.*, 1982): occasionally the thrusts are NW-dipping and SE-vergent (e.g. on Lane's Brook; Hyde, 1979).

INTERPRETATION

Three models have been proposed for the origin of the structures in the Late Paleozoic rocks: rift (e.g. Belt, 1968) and wrench/pull-apart (e.g. Webb, 1969; Keppie, 1982; Bradley, 1982; Bradley and Bradley, 1986), and a fold and thrust belt in a foreland or retro-arc basin (Keppie, 1982). Bradley (1982) inferred >200 km dextral motion on bounding NE-trending faults (Cabot and Bellisle/Lubec fault systems) and where they are offset the Magdalen Basin formed. He interpreted the stratigraphic record in terms of a two-stage process: (1) initial stretching and thinning of the lithosphere accompanied by rapid subsidence and volcanism during the late Devonian to Viséan, followed by (2) gradual thermal subsidence during which the depositional basins expanded in the late Carboniferous to Permian. However, Bradley (1982) acknowledges that the "steer's head" cross-section predicted by McKenzie's (1978) model is not developed. This does not eliminate the pull-apart model because both theoretical studies and observations of many small strike-slip basins show that the thermal anomaly decays during rifting (Pitman and Andrews, 1985; Aydin and Nur, 1982). Also, three episodes of volcanism, rather than one, are present.

Thus, the development of the Magdalen Basin and associated basins as pull-apart basins depends upon the recogition of other factors, such as geological mismatches, longitudinal and lateral basin asymmetry, episodic rapid subsidence, development of pronouced topographic relief, marked differences in stratigraphic thickness, facies geometry and the occurrence of unconformities from one basin to another, and strike-slip fault kinematics (Christie-Blick and Biddle, 1985). Geological mismatches have not been recorded. While basin asymmetry is evident in NW-SE cross-section of the Devonian-Tournaisian it generally has not been noted in NE-SW sections. Indeed, seismic sections across the Magdalen and associated basins show little evidence of basin asymmetry (Marillier *et al.*, 1989). While, four episodes of rapid subsidence have been recorded, they appear to be broadly synchronous in all the basins. Although, Webb (1969) inferred dextral movements on many NE-SW faults, many of these offsets are not unequivocal (discussion of Webb, 1969). In other cases the evidence for offset has been disproved.

For example, Webb (1969) and Belt (1969) inferred at least 100 km dextral displacement on the Cabot fault based upon the correlation of two mafic bodies cut by the fault: these bodies have since been shown to be of different ages (Knight, 1983). Any component of strike slip displacement on the Cabot fault must have been minor because detritus in rock units ranging in the age from (?)Late Devonian to Late Carboniferous has been traced to the distinctive alkaline Topsails Igneous Complex and adjacent units lying immediately southeast of the fault (Hyde, 1979; Knight, 1983). Horizontal, inclined and vertical slickensides have all been recorded indicating a complex fault history. Furthermore, dextral motions on NE-trending faults would produce an extensional regime around the St. Lawrence Promontory, and this is incompatible with the N-S contractional setting observed in the Gaspe Peninsula.

On the other hand, dextral movements on E-trending faults is well documented from late Devonian to Permian and these movements change into oblique thrusts where W-trending faults swing southwest. This suggests that the relative motion between Laurentia and outboard zones was east-west. In such a kinematic regime, minor dextral movement could be expected on NE-trending faults along with components of both normal and reverse movement depending upon local fault geometries (Crowell, 1974). The inherited NW-trend of the Appalachians around the St. Lawrence Promontory would induce an extensional setting in the Magdalen Basin (Fig. 18). Furthermore, the bend in the Appalachians through the Quebec Reentrant would be the locus of contraction and could cause a northerly swing in the principle stress direction.

The extent of the shortening which took place during the Devonian, Carboniferous and Permian is difficult to estimate. If the basal thrust beneath the Meguma Zone extends at least to the edge of the present continental slope (Fig. 1), a minimum east to west displacement of 600 km is indicated (this does not include any internal shortening). The occurrence of pebbles of Lower Devonian Torbrook Formation (containing an unique Rhenish fauna found in situ only in the Meguma Terrane) in conglomerates of the Tournaisian Horton Group in southern Cape Breton Island (Keppie, 1982) allow this relative displacement to be partitioned into Devonian and Permo-Carboniferous portions: 300 and 300 km (assuming that the Meguma Terrane collided with southern New Brunswick), respectively. This gives relative displacement rates of 0.6 cm/year and 0.3 cm/year, respectively. It is clear from the data presented earlier that Devonian-Permian overthrusting extended to the western margin of the Canadian Appalachians, however the amount of displacement is unknown.

Diachronism in the onset of Devonian deformation is indicated by the Early Devonian onset of deformation in the uppermost allochthon (the Meguma Zone) and the Late Devonian onset farther northwest on the Gaspe Peninsula. This diachronism may be related to progressive telescoping associated with accretion of the Meguma Zone to the southeastern margin of the Avalon Zone (Keppie and Dallmeyer, 1987; Dallmeyer and Keppie, 1987). Paleomagnetic reconstructions suggest that following collision of Gondwana (South America) with Laurentia during the Silurian, a major ocean opened between Gondwana and Laurentia during the Devonian (Van der Voo, 1988; Kent and van der Voo, 1990). The earliest Devonian deformation in the Meguma Zone suggests that its accretion was associated with the collision of South America and Laurentia. The continued deformation experienced by the Meguma Zone throughout the Devonian appears to incompatible with extension associated with a major Devonian ocean. An alternative model proposed by Keppie (1977) proposed anticlockwise rotation of Gondwana relative to Laurentia during the Devonian and would produce the observed dextral transpressive environment along the Appalachians. This would be followed, by oblique dextral continent-continent collision between Africa and Laurentia in the Carboniferous and Permian. This deformation appears to have synchronously affected the entire width of the Canadian Appalachians. This contrasts with the limited deformation along the subduction zone as occurs in ocean-ocean or ocean-continent convergence that is apparent in the Ordovician-Devonian deformation.

Depending upon the polarity of the subduction zone during the terminal collision, a major foreland basin on the subducting plate or a segmented retro-arc basin on the obducted plate would be formed beyond the mountain front. Both basins could eventually be involved in the fold and thrust belt deformation. An examination of the Appalachians shows that the major foreland basin in the southern Appalachians is replaced by fragmented Late Paleozoic basins in the northern Appalachians (Williams, 1978). The transition appears to coincide with the main oroclinal bend in the orogen near the Pennsylvania - New York State boundary at 41° latitude. This suggests a reversal of polarity of subduction at this latitude (Keppie, 1989). The style of deformation in the northern Appalachians, where thrusts swing into transcurrent faults resembles shovel-shaped thrusts or the stacking of shingles. The oblique E-W collision between Gondwana and North America in the NE-SW trending Canadian Appalachians would produce a component of dextral shear along the orogen. This could lead to oblique wrench faults parallel to the strike and pull-apart basins (such as the Magdalen Basin) at Z-bends in the orogen. Thus, within the gross retro-arc fold-and-thrust belt, many other smaller scale tectonic features may form. This would appear to explain the local complexity of the movements recorded in some areas.

MESOZOIC STRUCTURES

Mesozoic rocks are preserved mainly along the eastern seaboard of North America. Some Paleozoic structures were reactivated during the rifting stage of the opening of the Atlantic Ocean. The largest of these structures is the Bay of Fundy. Seismic data show the Bay of Fundy to be a half graben in which the listric normal fault along the southern coast of New Brunswick appears to have reactivated the late Paleozoic listric thrust passing beneath the Meguma Terrane (Keen *et al.*, 1991) (Fig. 1). Less translation took place in opening Chignecto Bay at the head of the Bay of Fundy, the difference being taken up along a sinistral, E-W flower structure (the Minas Fault System) passing from the Minas Basin into Chedabucto Bay. Associated with these movements, a series of NW-SE sinistral faults and kink bands cut the Meguma Terrane (Keppie, 1983). The largest of these faults is the Country Harbour Fault, that displaces the northern boundary of the Meguma Terrane. Late N-S and NW-SE faults in northern Nova Scotia and New Brunswick may also be of this age, as may be similar structures in Newfoundland.

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